Department of Natural Resources
Resource Assessment Service
MARYLAND GEOLOGICAL SURVEY
Emery T. Cleaves, Director

REPORT OF INVESTIGATIONS NO. 65

HYDROGEOLOGY AND SIMULATION OF GROUND-WATER FLOW IN THE UPPER WICOMICO RIVER BASIN AND ESTIMATION OF CONTRIBUTING AREAS OF THE CITY OF SALISBURY WELL FIELDS, WICOMICO COUNTY, MARYLAND

by

David C. Andreasen
and
Barry S. Smith

Prepared in cooperation with the
United States Department of the Interior
Geological Survey
and
The City of Salisbury Department of Public Works

1997

CONVERSION FACTORS AND VERTICAL DATUM

Multiply	by	To obtain
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
foot squared per day (ft^2/d)	0.09290	meter squared per day (m^2/d)
square mile (mi^2)	2.590	square kilometer (km^2)
gallon (gal)	3.785	liter (L)
foot per day (ft/d)	0.3048	meter per day (m/d)
inch per year (in./yr)	25.4	millimeter per year (mm/yr)
million gallons per day (Mgal/d)	0.04381	cubic meter per second (m^3/s)
cubic feet per day (ft^3/d)	0.02832	cubic meter per day (m^3/d)
gallon per day (gal/d)	0.003785	cubic meter per day (m^3/d)
gallon per minute (gal/min)	0.06309	liter per second (L/s)

Chemical concentrations are expressed in milligrams per liter (mg/L).

Temperature in degrees Celsius (C) can be converted to degrees Fahrenheit (F) using the following linear equation:

$$F = 1.8C + 32.$$

Sea level: In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929--a geodetic datum derived from a general adjustment of the first-order level nets of the United States and Canada, formerly called Sea Level Datum of 1929.

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ABSTRACT

The entire water supply for the City of Salisbury, Maryland, located in the drainage basin of the Upper Wicomico River on the lower Delmarva Peninsula, is provided by two well fields that tap a shallow, unconfined Coastal Plain aquifer (Salisbury aquifer). In 1993, average daily pumpage was 3.5 million gallons per day from the northern well field ("Paleochannel well field") and 1.7 million gallons per day from the southern well field ("Park well field"). Total pumpage from the Salisbury aquifer in the basin was approximately 8.7 million gallons per day in 1993. Industrial, residential, commercial, and agricultural land-use areas are located near the well fields. The aquifer is thus potentially vulnerable to both point-sources (such as disposal and storage sites) and non-point sources of pollution (such as pesticides and road salt). Data from hydrogeologic analyses were applied using the U.S. Geological Survey's three-dimensional ground-water-flow model (MODFLOW) and particle-tracking routine (MODPATH) to determine areas contributing recharge to the city's well fields under current and future pumping conditions and to identify possible sources of contamination. The study, conducted between July 1993 and June 1996, was a cooperative effort between the City of Salisbury Department of Public Works, the Maryland Geological Survey, and the U.S. Geological Survey.

The Salisbury aquifer in this area is composed of coarse sand and gravel and has a high transmissivity. Transmissivity values range between 8,200 feet squared per day and 57,000 feet squared per day with an average of about 22,000 feet squared per day in the Park well field to as much as 53,500 feet squared per day in the Paleochannel well field, where an unusually thick (greater that 200 feet) paleochannel deposit is present. In the paleochannel, the confining bed at the base of the Salisbury aquifer is breached, resulting in a hydraulic interconnection with the underlying Manokin aquifer.

The Salisbury aquifer receives recharge from precipitation in topographically high areas in the basin and discharges water to numerous small streams, man-made ponds, the tidal part of the Wicomico River, and public and private water-supply wells. Because the aquifer is unconfined, has a high permeability, and is recharged rapidly from percolating precipitation, water levels respond to seasonal changes in recharge but show only small local responses to pumping. The average net ground-water recharge rate is estimated at 13.4 inches per year from baseflow separation of long-term streamflow record in a sub-basin (Beaverdam Creek basin). The Salisbury aquifer shows good hydraulic connection to streams within the basin. Pumping from Paleochannel well 1 lowered baseflow by 0.1 cubic foot per second (approximately 60 percent of total stream flow) in Little Burnt Branch, a small tributary that passes near the well. Results of a three-dimensional, finite-difference ground-water flow model (MODFLOW) indicate that average outflow to the tidal parts of the Wicomico River and streams in the basin is approximately 4.7 and 44 million gallons per day, respectively.

Contributing areas of the city's wells extend upgradient from the wells approximately 1,000 feet for the 0- to 1-year time zone; 6,000 feet for the 1- to 10-year time zone, and 10,000 feet for the 20-year time zone. Most of the water entering the water table within these time periods is captured by the top part of the well screens. The areal patterns of the contributing areas are irregular and are affected by streams and ponds. The position and shape of the contributing areas vary with slight changes in hydraulic properties input to the ground-water-flow model. For example, a 7-percent increase in recharge caused the west-trending contributing area for Paleochannel well 1 to become narrower and to shift northward slightly. The northeast-trending

portion is approximately 500 feet longer than the contributing area based on the calibrated model. A 10-percent reduction and increase in aquifer porosity lengthens and shortens, respectively, the contributing areas for Paleochannel well 1 approximately by 500 feet. Contributing areas based on the estimated pumpage in 2010, which represents an increase of 15 percent from 1993 pumpage, show a general broadening of less than 500 feet for most of the city's wells. The contributing areas of the Paleochannel well field are predominantly in agricultural and open areas, whereas the contributing areas of the Park well field are predominantly in commercial and residential areas.

Water particles tracked forward from sites with the potential to be sources of contamination to the Salisbury aquifer indicate that recharge entering the aquifer at two sites--the Conrail railroad and a site with a ground-water discharge permit located at Route 13 and Naylor Mill Road--would be captured by Paleochannel well 2 after 2 and 11 years of travel time, respectively.

Recharge dates for water sampled at various depths in the Salisbury aquifer near the well fields, estimated using tritium and chlorofluorocarbon analysis, ranged from pre-1963 to 1992. The older water was withdrawn from the deeper wells and the younger water was withdrawn from the shallower wells. The oldest water sampled was from the deep part of the paleochannel where mixing with artesian Manokin aquifer water occurs. The range of recharge ages determined by use of tritium and chlorofluorocarbons generally agree with the median recharge travel times calculated by the flow model.

INTRODUCTION

BACKGROUND

The City of Salisbury, in central Wicomico County, Maryland, is the commercial hub of the lower Delmarva Peninsula. The city is in the Upper Wicomico River drainage basin, which includes two major tributaries--the North Prong Wicomico River and Beaverdam Creek (fig. 1). Salisbury depends entirely on shallow ground water from the Salisbury aquifer for its water supply. The Salisbury aquifer is an attractive source of ground water because of its generally good water quality, high supply capacity, and shallow depths required for well construction. Also, deeper aquifers in the area tend to have high concentrations of dissolved solids, including high chloride levels.

Two well fields supply water to the city from depths of 60 to 180 feet (ft) below land surface. The older of the two well fields (Park well field) was established in 1925 in a city park along a dammed section of the South Prong Wicomico River (also referred to as Park Pond) near the intersection of Routes 50 and 13 (figs. 1 and 2). The Park well field consists of nine active production wells, two of which (WI Cf 201 and 202) were drilled in 1994. The second well field (Paleochannel well field) was established in 1967 with the construction of well 1 (WI Ce 200) in the highly productive paleochannel sediments north of Salisbury along Naylor Mill Road (figs. 1 and 2). A second production well (WI Ce 241) was completed in this field in 1989. Each well field has its own water-treatment plant.

Both well fields pump from the Salisbury aquifer, a water-table aquifer consisting of relatively thick, pebbly sand. The Salisbury aquifer is very productive with transmissivities reported between 8,200 feet per day (ft^2/d) and 57,000 ft^2/d in the Park well field (Rasmussen and Slaughter, 1955) and up to 53,500 ft^2/d in the Paleochannel well field (Boggess and Heidel, 1968; Mack and Thomas, 1972). Daily pumpage in 1993 averaged approximately 1.7 million gallons per day (Mgal/d) in the Park well field and 3.5 Mgal/d in the Paleochannel well field. Peak pumpage occurred in July with a combined average of 6.1 Mgal/d pumped from the two well fields.

In 1991, modifications in the distribution system were made to release water from the new (Paleochannel) well field to the service area of the old well field. This interconnection, plus the construction of the second production well in the Paleochannel well field, has addressed the city's water-supply needs for the foreseeable future.

The Salisbury aquifer is susceptible to contamination because it is unconfined and has a high transmissivity. It is potentially vulnerable to contaminants from both point-sources (such as disposal and storage sites or buried tanks) and non-point sources of pollution (such as pesticides and road salt).

Pumpage from the city's well fields creates cones of depression in the water-table surface, which capture recharge from precipitation and infiltrating river water. To ensure the continued good quality of Salisbury's water supply, it is imperative to determine the areal extent of the contributing areas for the city's well fields and the rates of ground-water movement toward the pumping wells.

This study, conducted between July 1993 and June 1996, was a cooperative effort between the City of Salisbury Department of Public Works, the Maryland Geological Survey, and the U.S. Geological Survey.

PURPOSE AND SCOPE

The purpose of this report is to refine the hydrogeology of the Salisbury area and to determine the ground-water contributing areas of wells in the City of Salisbury's two well fields. This information is intended to be used by the city to develop a wellhead protection strategy designed to protect the water supply from contamination.

The first part of this report describes the hydrogeology of the Salisbury area, including the hydrogeologic framework, head distribution and direction of ground-water flow, ground-water/surface-water interactions, ground-water budget, and ground-water age. The middle section of this report describes a ground-water-flow model that simulates ground-water flow in the Salisbury and Manokin aquifers in the Upper Wicomico River Basin. A discussion of the hydrologic-budget components used in the model, model calibration, and sensitivity analysis is included. The last part of this report presents the results of a particle-tracking simulation used to define the contributing areas of the city's two well fields and the direction of ground-water flow from potential contaminant sites.

LOCATION OF STUDY AREA AND PHYSICAL SETTING

The study area is on Maryland's Eastern Shore in the Upper Wicomico River Basin in central Wicomico County (fig. 1). Although the majority of the basin lies within Maryland's border, a small area of about 2 square miles (mi^2) in the northern part of the basin falls in Sussex County, Delaware (fig. 1). To evaluate the ground-water flow system, the entire Upper Wicomico River drainage basin, an area of roughly 80 mi^2, was included in the study area.

The City of Salisbury is located in the southwestern part of the basin. The areas of primary concern to this study are the two city well fields--the Paleochannel well field and the Park well field. The Paleochannel well field is in a pri-

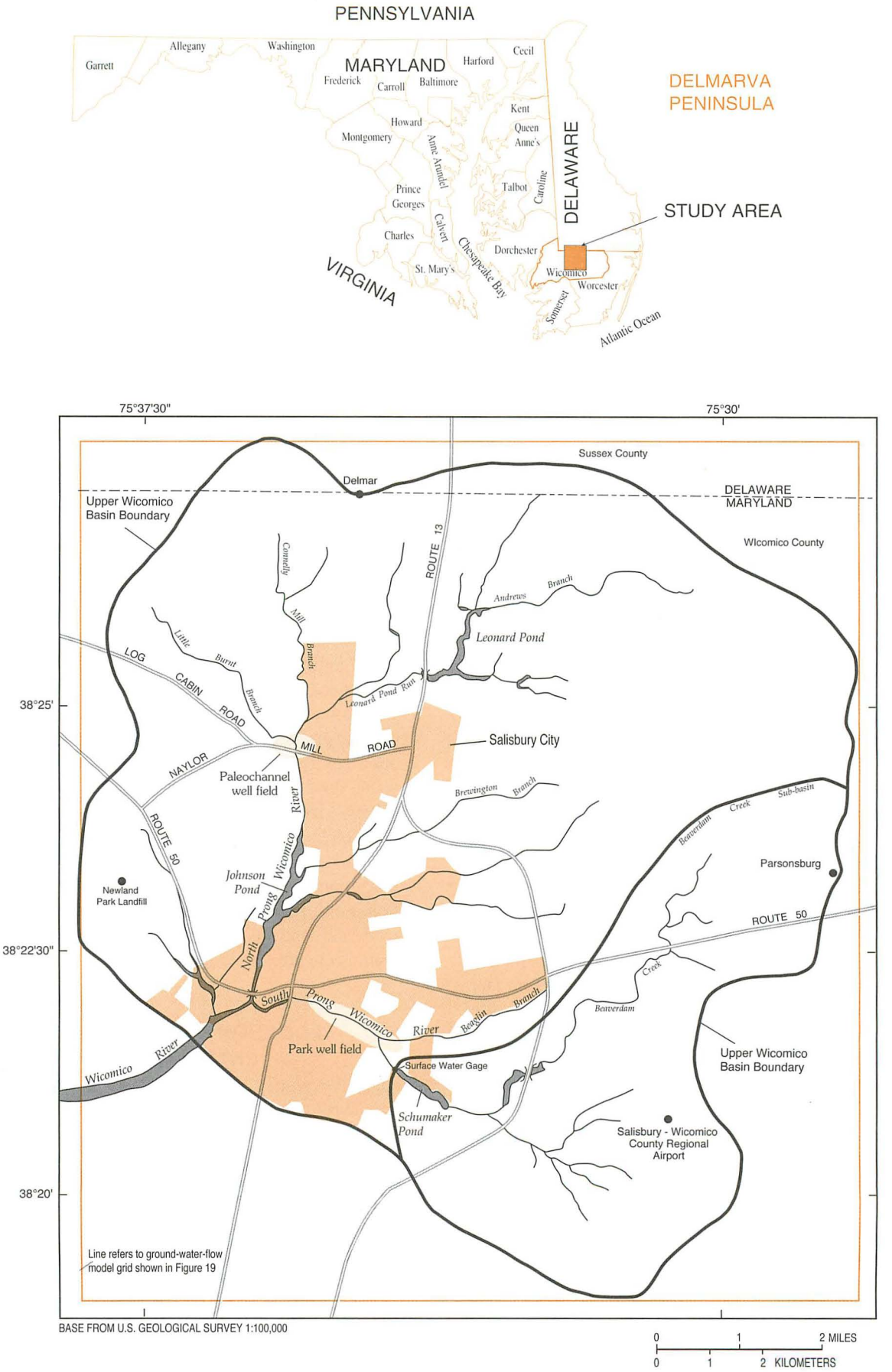


Figure 1 - Location of Study Area

marily rural and agricultural area. An industrial park is located approximately 0.5 mi to the east of the well field. The Park well field is located in a park in the center of the city. The park runs along a pond bordered by residential and light-commercial areas.

The study area lies within the Atlantic Coastal Plain Physiographic Province. This area is characterized by low topographic relief. Altitudes range from sea level to about 80 ft above sea level. The area is dissected by numerous small streams and irrigation ditches that feed two larger streams--the North Prong Wicomico River and Beaverdam Creek. Both streams are classified as third-order streams. The area is generally covered by very permeable coarse-grained sand that readily absorbs precipitation. Clayey soils cover the sand in some places.

Land use in the northern part of the study area is mostly agricultural. Land use in the south is a mixture of residential, agricultural, and commercial. Five major ponds, once used to power grain mills, are now used primarily for recreation and flood control in the basin. Numerous smaller ponds are also used for irrigation.

METHODS OF INVESTIGATION

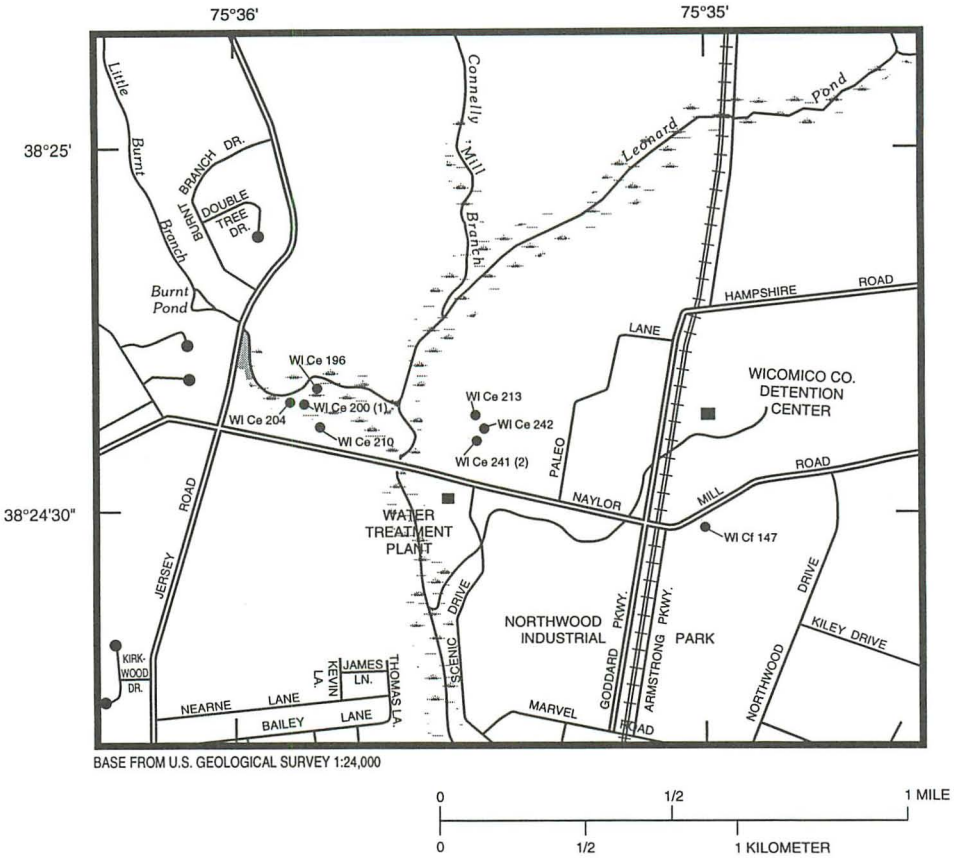
Information on the hydrogeology of the Salisbury area, including data on aquifer and confining bed characteristics, pumpage, head, and baseflow, were obtained mostly from literature and unpublished data on file at the Maryland Geological Survey and U.S. Geological Survey. Water levels were measured in 28 wells during the fall of 1993 and spring of 1994. Selected well records are presented in appendix A and the locations of the wells are plotted on a map (pl. 1). Baseflow seepage measurements were made during the fall of 1993 and 1994 along Little Burnt Branch by personnel from the U.S. Geological Survey, Dover, Delaware. Head measurements and water sample collection in the stream bottom of Little Burnt Branch were done using a potentiomanometer during the 1994 seepage run. The lithologic character of the sediments underlying Park Pond adjacent to the Park well field was examined in hand-driven, split-spoon core samples. Descriptions of the lithology of the sediments are given in appendix B. These data were used to construct structure-contour maps, a water-table map, hydrogeologic cross sections, water-level hydrographs, and data sets for input into, and calibration of, a digital, three-dimensional ground-water-flow model. The program used for ground-water-flow simulation was the U.S. Geological Survey's MODFLOW code (McDonald and Harbaugh, 1988). A second computer program--U.S. Geological Survey's MODPATH code (Pollock, 1989)--was used to track water particles through time to determine areas contributing water to the city's wells and to estimate the direction of ground-water flow from potential contaminant sites.

Wells are identified in this report using the City of Salisbury's well numbers (such as Paleochannel well 1) and the Maryland Geological Survey well numbers (such as WI Ce 200). In the Maryland Geological Survey's well-numbering system, the first two letters are the county prefix (for example, WI for Wicomico County). The second part of the well number consists of two letters that designate a 5-minute quadrangle within the county; the first letter (a capital letter) denotes a 5-minute segment of longitude from north to south, and the second letter (lower case) denotes a 5-minute segment of longitude from west to east. The location of the 5-minute quadrangles in the study are shown on plate 1. The wells are numbered sequentially in the order they were inventoried within each 5-minute quadrangle.

Water-quality data were collected from eight wells during the fall of 1993 for the purpose of age-dating the ground water. Tritium samples from eight wells were analyzed by the University of Waterloo, Ontario, Canada. Samples for tritium analysis were collected in 1 liter (L) glass bottles. Chlorofluorocarbons from seven wells were analyzed by Eurybiades Busenberg of the U.S. Geological Survey, Reston, Virginia. Collection of samples for chlorofluorocarbon analysis consisted of (1) purging the wells of water stored in the casing, (2) evacuating water using a stainless-steel submersible gas-reciprocating pump and 0.25 inch (in.) copper discharge line, (3) filling 62 milliliter (mL) borosilicate ampules under a nitrogen atmosphere, and (4) welding the ampules shut. Six samples were collected at each well. Tritium samples were collected midway through the chlorofluorocarbon sampling. Dissolved chloride and dissolved nitrate plus nitrite (as nitrogen)--possible indicators of young ground water--were analyzed in water taken from the same wells sampled for tritium and chlorofluorocarbon. The dissolved chloride and dissolved nitrate plus nitrite (as nitrogen) analyses were made by the U.S. Geological Survey National Water-Quality Laboratory in Denver, Colorado.

PREVIOUS INVESTIGATIONS

The first comprehensive study of the geology and ground-water resources of the Salisbury area was presented by Rasmussen and Slaughter (1955) in a tri-county report that included Somerset, Wicomico, and Worcester Counties. Incorporated in that study was a section by Meyer and Bennett (1955) on the Salisbury area that gave detailed descriptions of pumping tests at the city's Park well field. A detailed investigation into the hydrologic budget of the Beaverdam Creek basin, a subbasin within this study area, was conducted by Rasmussen and Andreasen (1959). Hansen (1966) described the stratigraphy of the Salisbury area and developed a stratigraphic model for the development of the Naylor Mill paleochannel. A later investigation by Owens and Denny (1979b) also discussed the geology of the paleochannel as part of a broader study of the Upper Cenozoic deposits in the central part of the Delmarva Peninsula. A geologic map of Wicomico County was produced in 1979 at the scale 1:62,500 (Owens and Denny, 1979a). The quan-



PALEOCHANNEL WELL FIELD

EXPLANATION

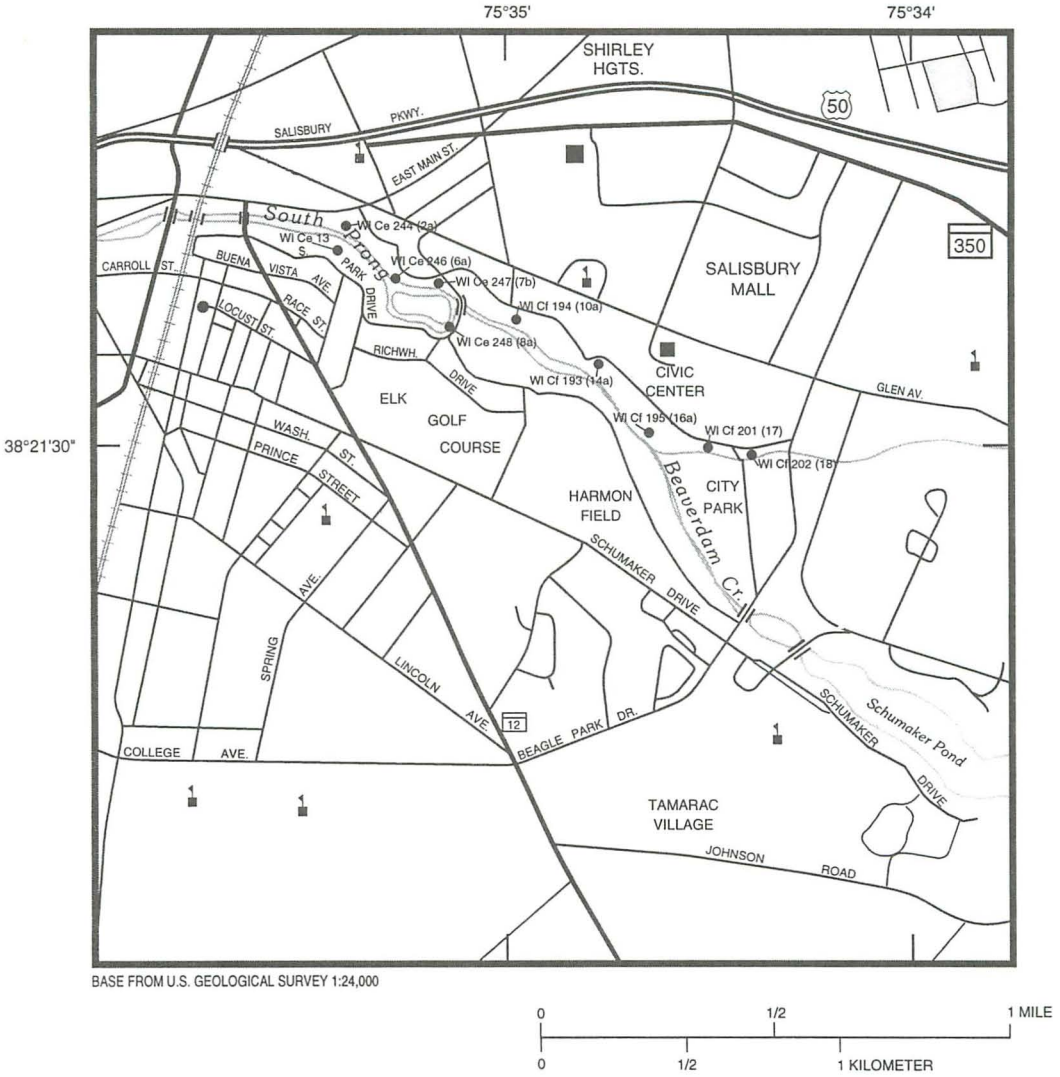
WI Ce 244 (2a) — WELL AND WELL NUMBER. NUMBER IN PARENTHESIS IS CITY OF SALISBURY'S WELL DESIGNATION AND INDICATES PUMPING WELL. ALL OTHER WELLS SHOWN ARE OBSERVATION WELLS.

Figure 2. - Location of selected wells in the Paleochannel and Park well fields.

tity and quality of the water resources locally available to the Salisbury area were investigated by Boggess and Heidel (1968). That study concentrated on the hydrogeology of the water-table aquifer and its relation to the paleochannel deposits and to surface water. A study to further define the extent and productivity of the paleochannel along Naylor Mill Road was completed in 1972 by Mack and Thomas, and by Weigle (1972). During the study by Mack and Thomas, a 30-day aquifer test was conducted to determine the hydraulic characteristics and water-supply potential of the paleochannel. Many observation wells in and around the Naylor Mill Road area were constructed at that time.

ACKNOWLEDGMENTS

The authors would like to express gratitude for the cooperation and help provided by the personnel of the City of



PARK WELL FIELD

Figure 2. - Continued.

Salisbury Department of Public Works and, in particular, Thomas D. Plotts (former Director), Wilmer Elliot (Director), and Newell W. Messick III (Deputy Director). Technical support was provided by William B. Fleck of the U.S. Geological Survey, Towson, Maryland, whose expertise in hydrogeology and ground-water-flow modeling greatly enhanced this report. Harry Hansen, of the Maryland Geological Survey, provided helpful expertise regarding the stratigraphy of the Salisbury area. Analysis and preliminary interpretation of the chlorofluorocarbon data were provided by Eurybiades Busenberg of the U.S. Geological Survey, Reston, Virginia. Robert Simmons of the U.S. Geological Survey, Dover, Delaware, provided necessary support for the stream baseflow measurements.

HYDROGEOLOGY

HYDROGEOLOGIC FRAMEWORK OF THE SALISBURY AREA

The Salisbury area is located within the Atlantic Coastal Plain Physiographic Province in Maryland. The Coastal Plain is composed predominantly of stratified sands, silts, and clays that range in age from Cretaceous to Pleistocene. The thickness of the sediments increases eastward from the western boundary with the Piedmont Province (Fall Line) to over 7,700 ft at Ocean City, Maryland (Hansen and Edwards, 1986). The sediments dip gently towards the southeast, forming wedge-shaped deposits that overlie an older, consolidated basement complex of Jurassic(?) to Precambrian age.

Most of the sedimentary wedge consists of non-marine and marginal-marine deposits of Cretaceous age. Overlying the Cretaceous units are marine, estuarine, and fluvial sediments of Tertiary age and mostly fluvial sediments of Quaternary age.

Deposits described in this report range in age from Tertiary to Quaternary. Geologic units include the St. Marys and Manokin Formations of Miocene age; Pensauken Formation, Beaverdam Sand, and Walston Silt of Miocene or Pliocene(?) age; and the Parsonsburg Sand of Pleistocene age (tab. 1). Hydrogeologic cross sections A-A' and B-B', trending east-west and north-south, respectively, are used to describe the structure of the various units discussed in the report (figs. 3 and 4).

The St. Marys Formation, a member of the Chesapeake Group, is a laterally extensive unit of marine origin. The formation consists predominantly of clayey silt, silty clay, very fine sand, and shell material (Rasmussen and Slaughter, 1955). The top of the formation lies about 170 ft below sea level at the west-central side of the Wicomico River Basin and slopes down to about 270 ft below sea level at Parsonsburg (figs. 3 and 5). Its thickness ranges from 93 ft at well WI Cf 77 in the Park well field to 77 ft at Delmar (well RD 31-13), and 30 feet at Hebron (Rasmussen and Slaughter, 1955). The St. Marys Formation functions as a confining layer below the aquifers included in this study.

Rasmussen and Slaughter (1955) assigned the Upper Miocene beds overlying the St. Marys Formation to the Yorktown and Cohansey(?) Formation, a marginal-marine to marine deposit consisting of two sand units and two clay units. Later stratigraphic studies showed, however, that in Maryland these beds correlate with neither the Yorktown Formation nor the Cohansey Formation. For this reason Hansen (1981), Andres (1986), and Pazzaglia (1993) suggested that the names be abandoned. Generally, in this report the beds between the St. Marys Formation and the Pensauken Formation are described using hydrogeologic names, which are retained from Rasmussen and Slaughter (1955). From youngest to oldest these are: the upper confining bed (clay), the Pocomoke aquifer (sand), the lower confining bed (clay), and the Manokin aquifer (sand). In the Salisbury area, the upper confining bed and most of the Pocomoke aquifer have been eroded leaving only the lower confining bed and the Manokin aquifer (Rasmussen and Slaughter, 1955). Owens and Denny (1979b) referred to the Manokin aquifer informally as the "Manokin bed." Andres (1986) raised the Manokin to formation rank and used that designation in Delaware, which is followed in this report. The Pocomoke aquifer is present only in the extreme eastern part of the Wicomico River Basin where it is hydraulically interconnected with the overlying Salisbury aquifer (Boggess and Heidel, 1968). The lower confining bed thickens to the southeast from about 50 feet in the central part of the Wicomico River Basin to about 70 feet at Parsonsburg (fig. 3).

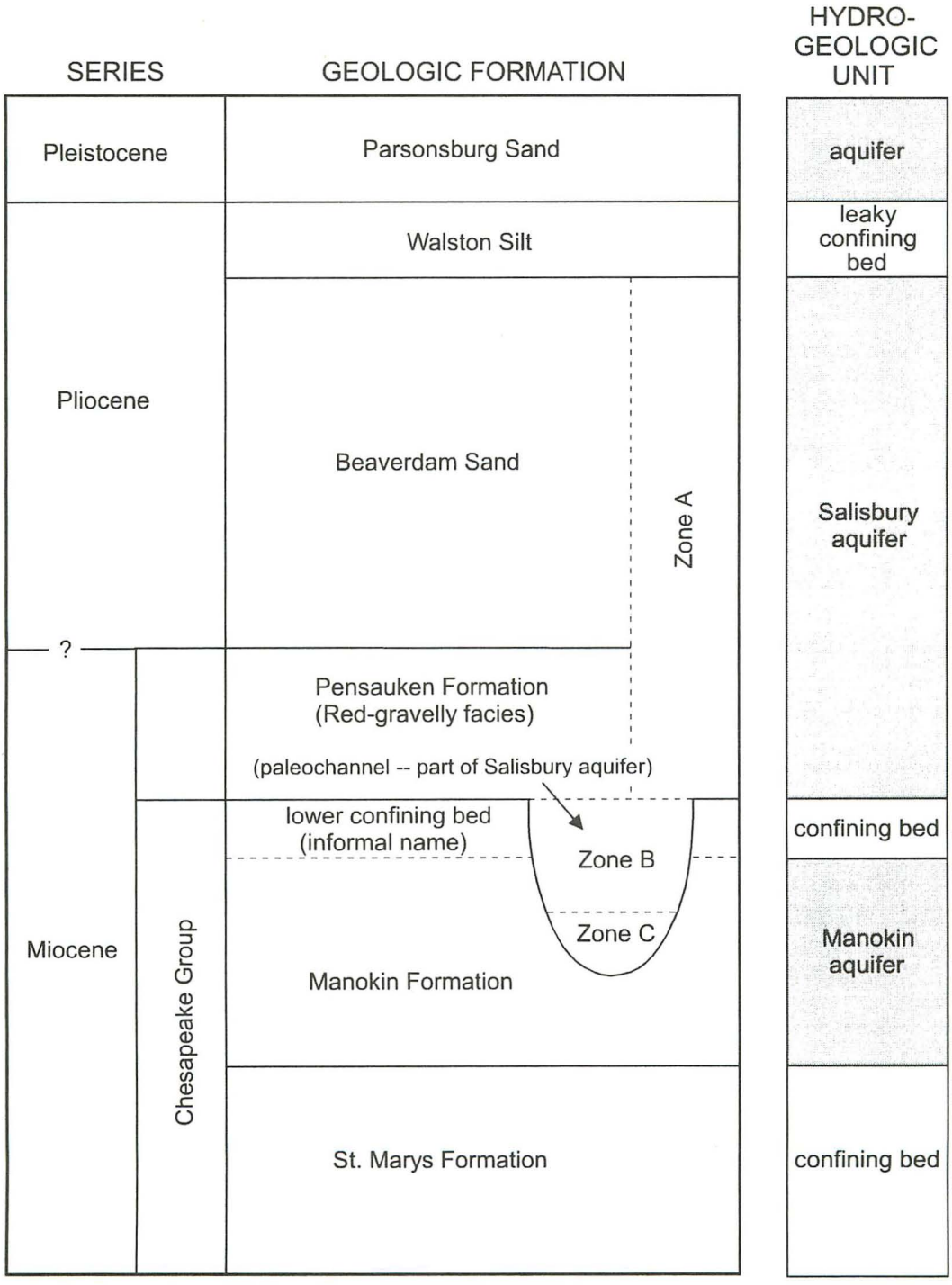
Owens and Denny (1979a) describe the Manokin Formation (same unit as Manokin aquifer) as dark gray, very clayey and silty, medium-grained sand, interbedded locally with fine gravel. In the subsurface, the Manokin aquifer appears on geophysical logs as predominantly a medium- to coarse-grained sand. The altitude of the top of the aquifer ranges from about 60 ft below sea level in the western part of the Wicomico River Basin to about 140 ft below sea level in the eastern part (fig. 5). The thickness of the Manokin aquifer ranges from 130 ft at Parsonsburg to about 85 ft near Allen (figs. 3 and 4). The Manokin aquifer is in direct contact with the overlying Pensauken Formation (at the base of the Salisbury aquifer) northwest of the Salisbury area and in areas where Pliocene(?)-age channels have completely eroded the lower confining bed.

The lower confining bed occurs throughout the study area outside of the paleochannel. The confining bed material consists of bluish-gray clay (Rasmussen and Slaughter, 1955). The thickness of the confining bed is approximately 25 ft in most areas and increases to about 65 ft in the eastern part of the basin (figs. 3 and 4).

The most distinctive subsurface feature in the Salisbury area is a paleochannel system cut into the lower confining bed and the Manokin aquifer (figs. 3 and 4). The main paleochannel trends southeastward from the northwest corner of Wicomico County to just north of Salisbury (Weigle, 1972). A smaller channel extends northward from that location and may reach Delmar. The main paleochannel cuts into the lower confining bed and Manokin aquifer approximately 180 ft below sea level and is 1.8 mi wide on the western side of the Wicomico River. The paleochannel narrows in the vicinity of Naylor Mill Road to less than 0.5 mi wide and deepens to about 200 ft below sea level. The smaller

(Text continued on p. 14.)

Table 1. - Tertiary-age stratigraphic and hydrogeologic units in the study area



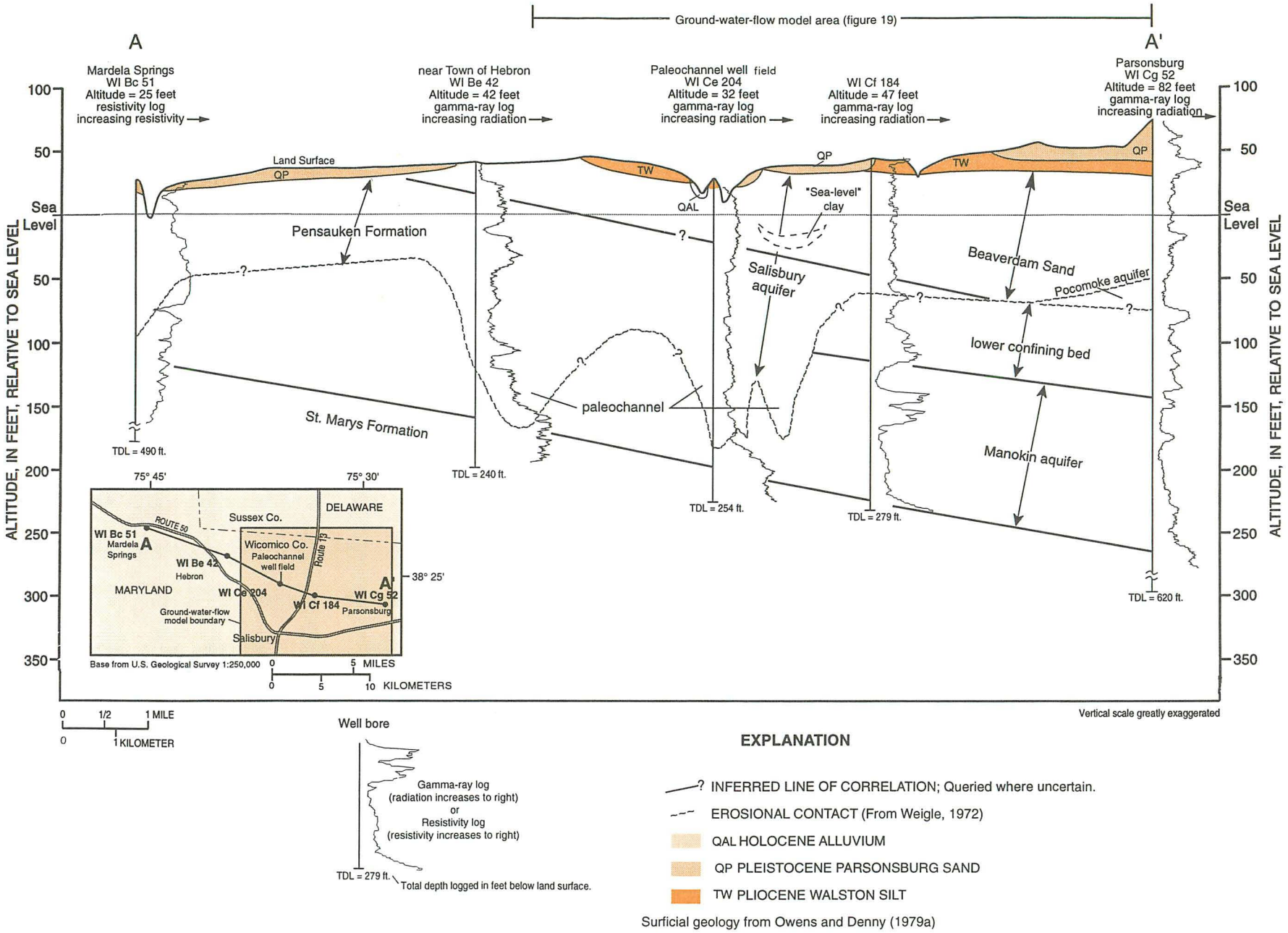


Figure 3. - Geohydrologic cross-section A-A', from Mardela Springs to Parsonsburg.

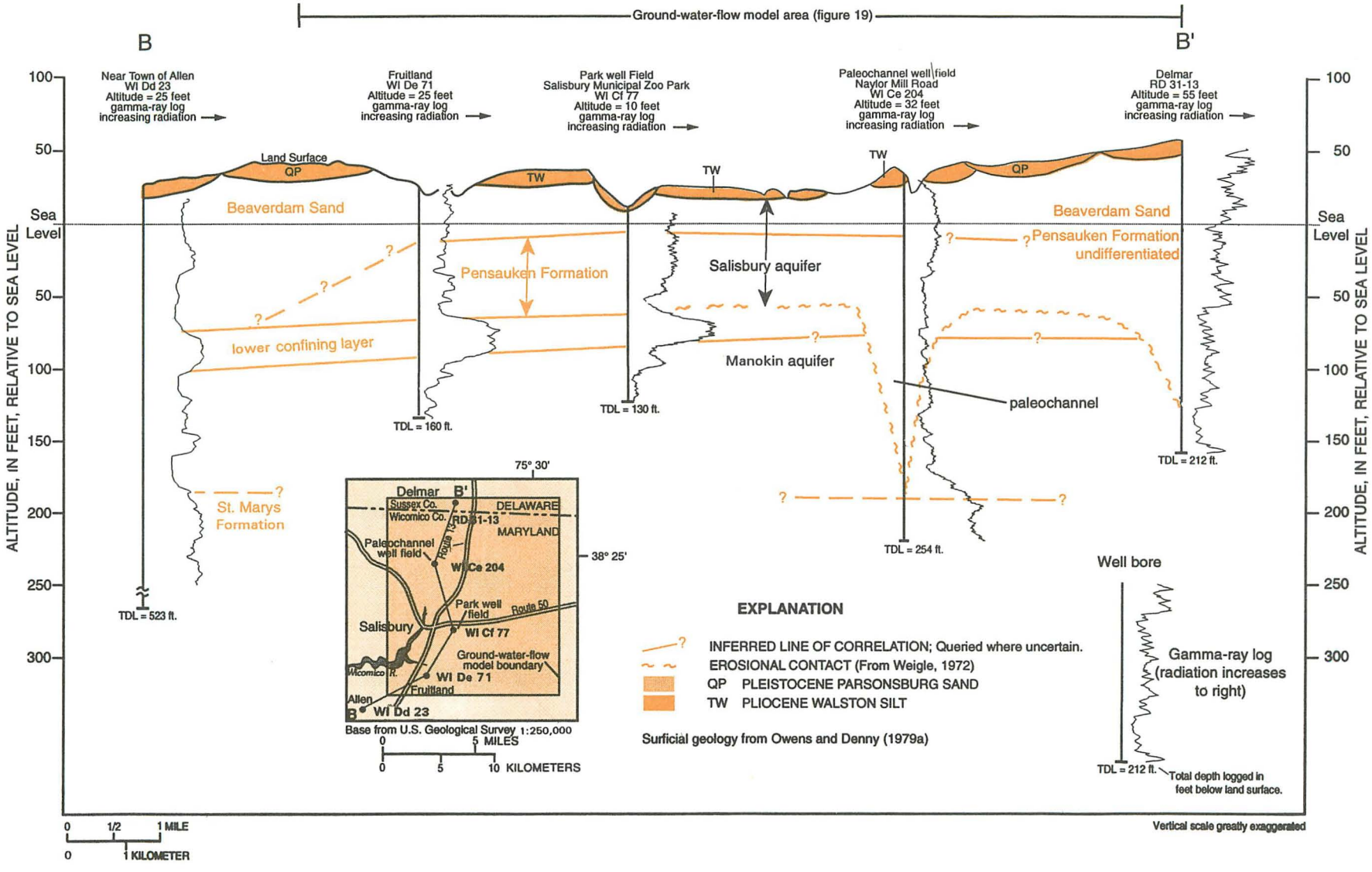
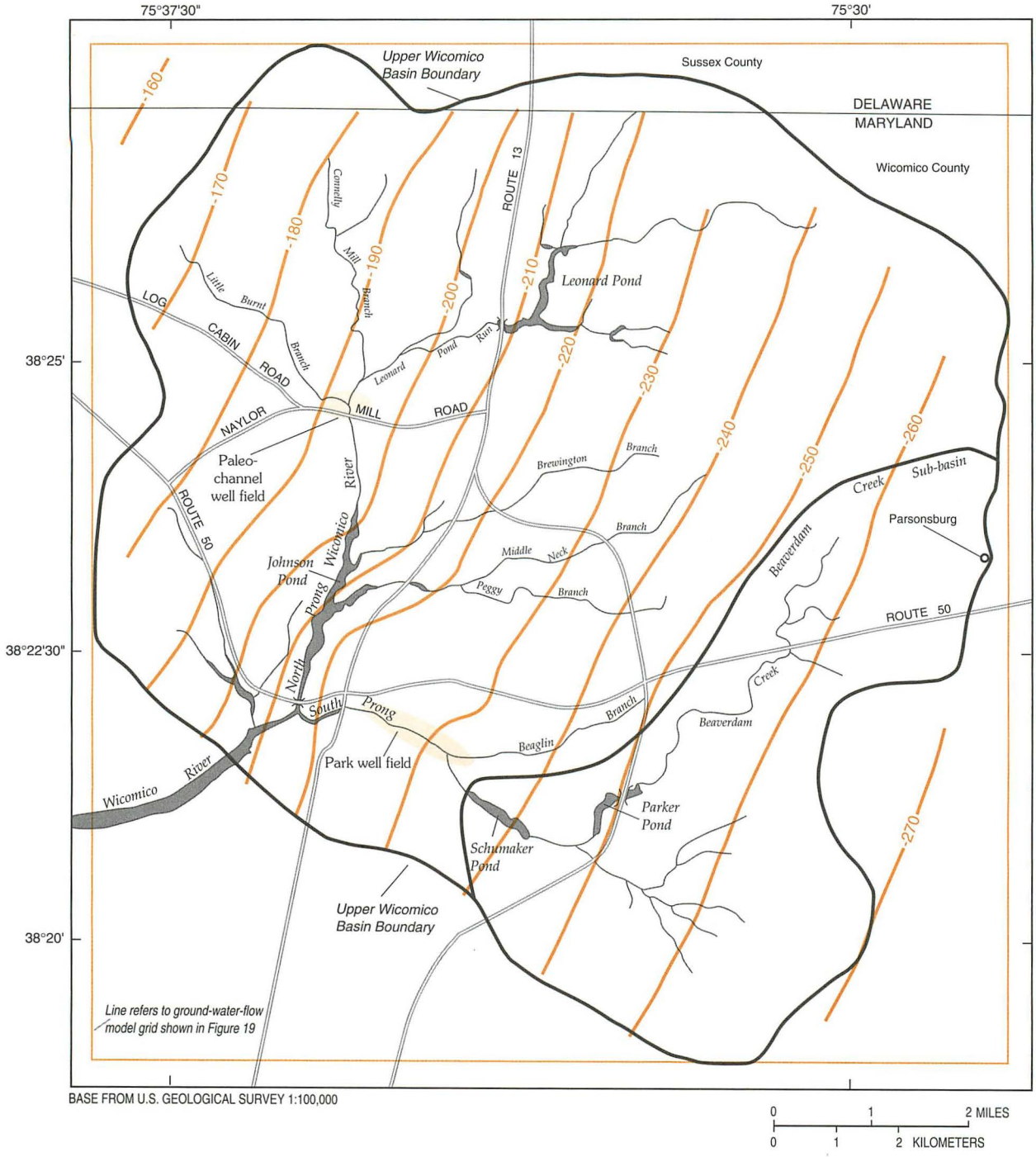


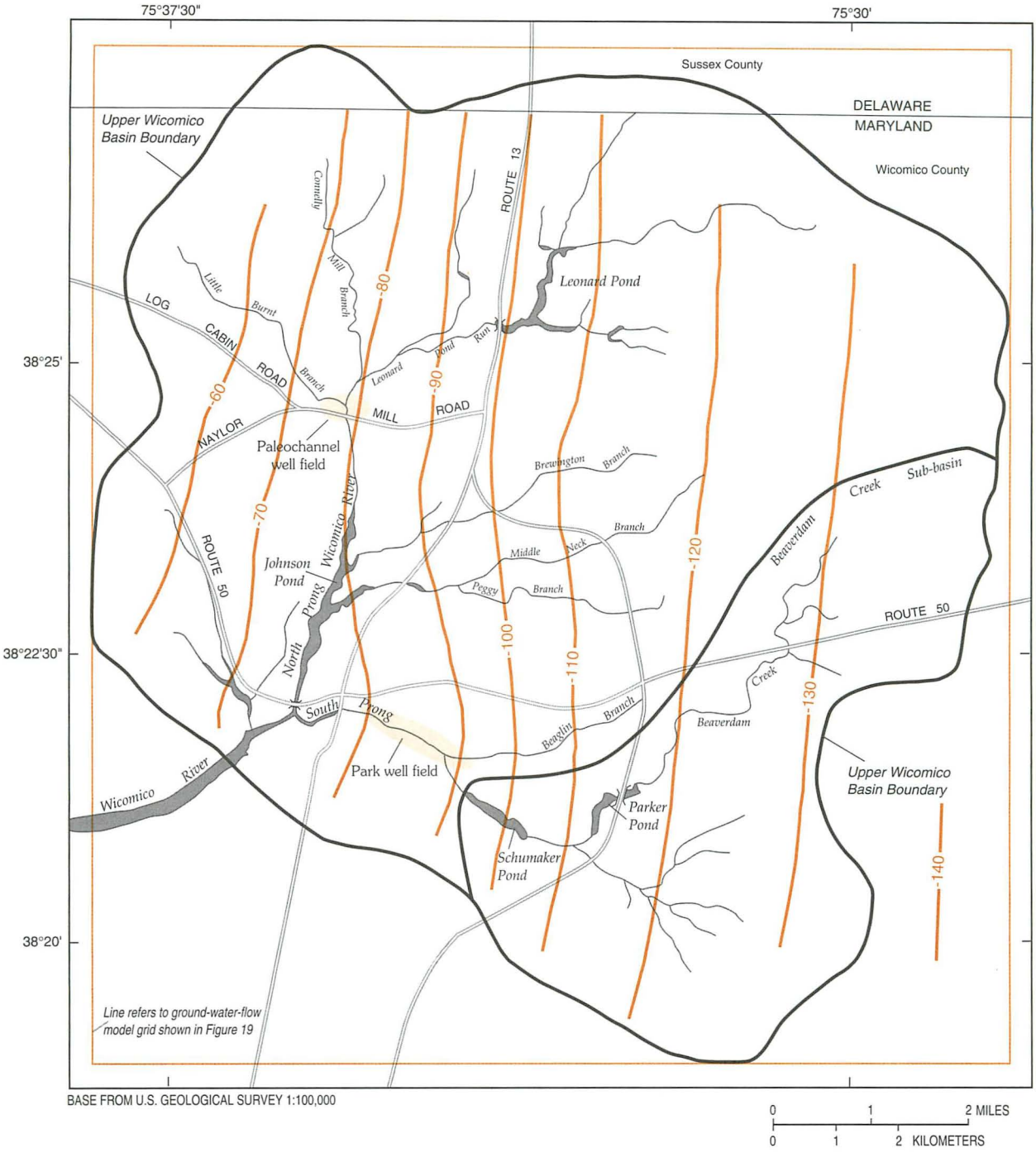
Figure 4. - Geohydrologic cross section B-B', from near Allen to Delmar.



EXPLANATION

-200 LINE OF EQUAL ALTITUDE, IN FEET RELATIVE TO SEA LEVEL. CONTOUR INTERVAL IS TEN FEET.

Figure 5. - Altitude of the top of the St. Marys Formation.



EXPLANATION

-100 LINE OF EQUAL ALTITUDE, IN FEET RELATIVE TO SEA LEVEL. CONTOUR INTERVAL IS TEN FEET.

Figure 6. - Altitude of the top of the Manokin aquifer.

paleochannel is approximately 0.6 mi wide and cuts through the lower confining bed and into the Manokin aquifer to a depth of approximately 90 ft below sea level at Naylor Mill Road (fig. 4). The structure of the paleochannel at Delmar is unclear. Weigle (1972) estimated the depth of the channel at 107 ft below sea level; however, some drillers' logs in the area show "blue" clay at a depth of 50 ft below sea level--a lithology more indicative of the Chesapeake Group than the paleochannel sediments.

The Pensauken Formation is described as a fluvial deposit laid down as a series of eastward-migrating channels (Owens and Denny, 1979b). Owens and Denny (1979a) suggest a late Miocene age for the Pensauken, based on pollen dating. Previously, beds of the Pensauken were thought to be of Plio-Pleistocene age and probably of fluvial or glacial outwash origin (Hansen, 1966). Owens and Denny (1979a) viewed the Pensauken as a fluvial facies of Rasmussen and Slaughter's (1955) Yorktown and Cohansey(?) Formation. This idea was rejected by Hansen (1981), in part because of the disconformable contact between the Pensauken and the Manokin Formation in the paleochannel area located north of Salisbury. The Pensauken appears to pinch out or change facies east of Salisbury.

The Pensauken Formation consists of orange to reddish-brown gravelly sand. This unit was named the "red gravelly facies" by Rasmussen and Slaughter (1955) based on the red coloration caused by oxidized iron in the sediment. In the area of the Naylor Mill paleochannel, the Pensauken is described as a tan, very coarse sand and gravel (Mack and Thomas, 1972).

An elongate, lenticular-shaped clay layer, referred to as the "Sea-level clay" by Mack and Thomas (1972) and Weigle (1972), occurs at a depth near sea level between wells WI Ce 204 and WI Cf 184 (fig. 3). In this report, as in Mack and Thomas (1972), the "Sea-level clay" is placed in the Pensauken. The clay layer is approximately 3.5 mi long and oriented in a northeast-southwest direction. The pink-brown and light-gray clay ranges up to 20 ft in thickness. Because the vertical movement of recharge water is lowered by the clay, the Salisbury aquifer may be less productive in this area.

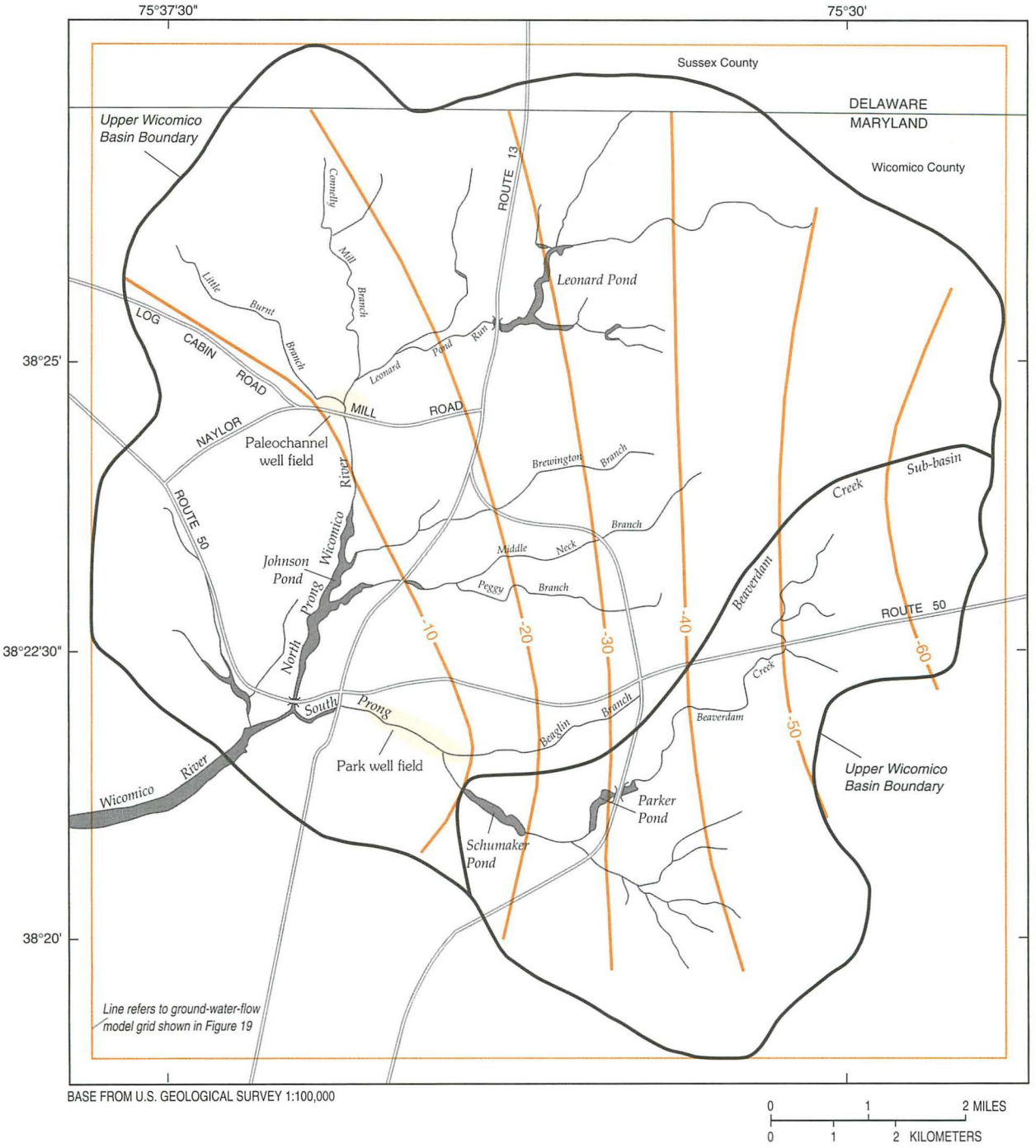
The Beaverdam Sand overlies the Pensauken Formation. It is a light-colored gravelly sand. The base of the Beaverdam Sand ranges in altitude from less than 10 ft below sea level in the vicinity of the Park well field to about 60 ft below sea level along the eastern edge of the basin (fig. 7). Owens and Denny (1979a) determined the age of the Beaverdam Sand to be Pliocene, based on its microflora fossils. The Beaverdam Sand overlies Upper Miocene deposits to the east of the Salisbury area and to the south near Allen as well (figs. 3 and 4). Hansen (1966) viewed the Beaverdam Sand as an estuarine unit deposited during an interglacial marine transgression. For this reason, Hansen (1966) considered the Beaverdam to be Pleistocene in age. However, more recent studies indicate that the onset of glacial-interglacial alternation occurred as early as the late Pliocene (Zimmerman and others, 1984).

The unconfined aquifer modeled in this report is the Salisbury aquifer (Hansen, 1966). It includes both the sands and gravels of the Pensauken Formation (formerly called the "red gravelly facies") and the Beaverdam Sand (tab. 1). The base of the aquifer outside the paleochannel area ranges in altitude from about 40 ft below sea level along the western edge of the basin to about 70 ft below sea level near the eastern edge (fig. 8). In the vicinity of the Paleochannel well field, the base of the aquifer occurs at about 175 ft below sea level. Boggess and Heidel (1968) mapped the saturated thickness of the Salisbury aquifer. Their map shows saturated thicknesses ranging from 40 to 60 ft in the topographically low areas, 80 to 100 ft in topographically high areas, and 100 to 200 ft in the paleochannel area along Naylor Mill Road and at Delmar. The saturated thickness of the Salisbury aquifer in 1993 is similar to that mapped by Boggess and Heidel, since the water-table surface has not changed significantly since that time. The Salisbury aquifer is the local equivalent to the Columbia aquifer. The Columbia aquifer consists of a widespread blanket of surficial sands and gravels which covers much of the upland areas of the Delmarva Peninsula (Bachman and Wilson, 1984).

The Walston Silt of Pliocene age overlies the Beaverdam Sand. It consists of dark-gray, pale gray or pale green clayey silt interbedded with thin layers of light-colored, poorly sorted medium sand (Owens and Denny, 1979b). This formation, of probable marine origin, underlies most of the Upper Wicomico River Basin (figs. 3 and 4). Stream channels cut through the Walston Silt into the Beaverdam Sand at many locations. The thickness of the Walston Silt within the study area varies considerably (a few feet up to 80 ft) and is thickest in the highlands east and southeast of Salisbury (Owens and Denny, 1979a).

The youngest recognized formation within the Wicomico River Basin is the Parsonsburg Sand of Pleistocene age. This formation consists of loose, medium- to coarse-grained sand. The Parsonsburg Sand is present mainly in the northeastern part of the Wicomico River Basin where it overlies the Walston Silt, and in the central section of the basin, where it truncates the Walston Silt and is in contact with the Beaverdam Sand (Owens and Denny, 1979a). In the Upper Wicomico Basin area, the Parsonsburg Sand and the Beaverdam Sand are hydraulically interconnected and part of the Salisbury aquifer.

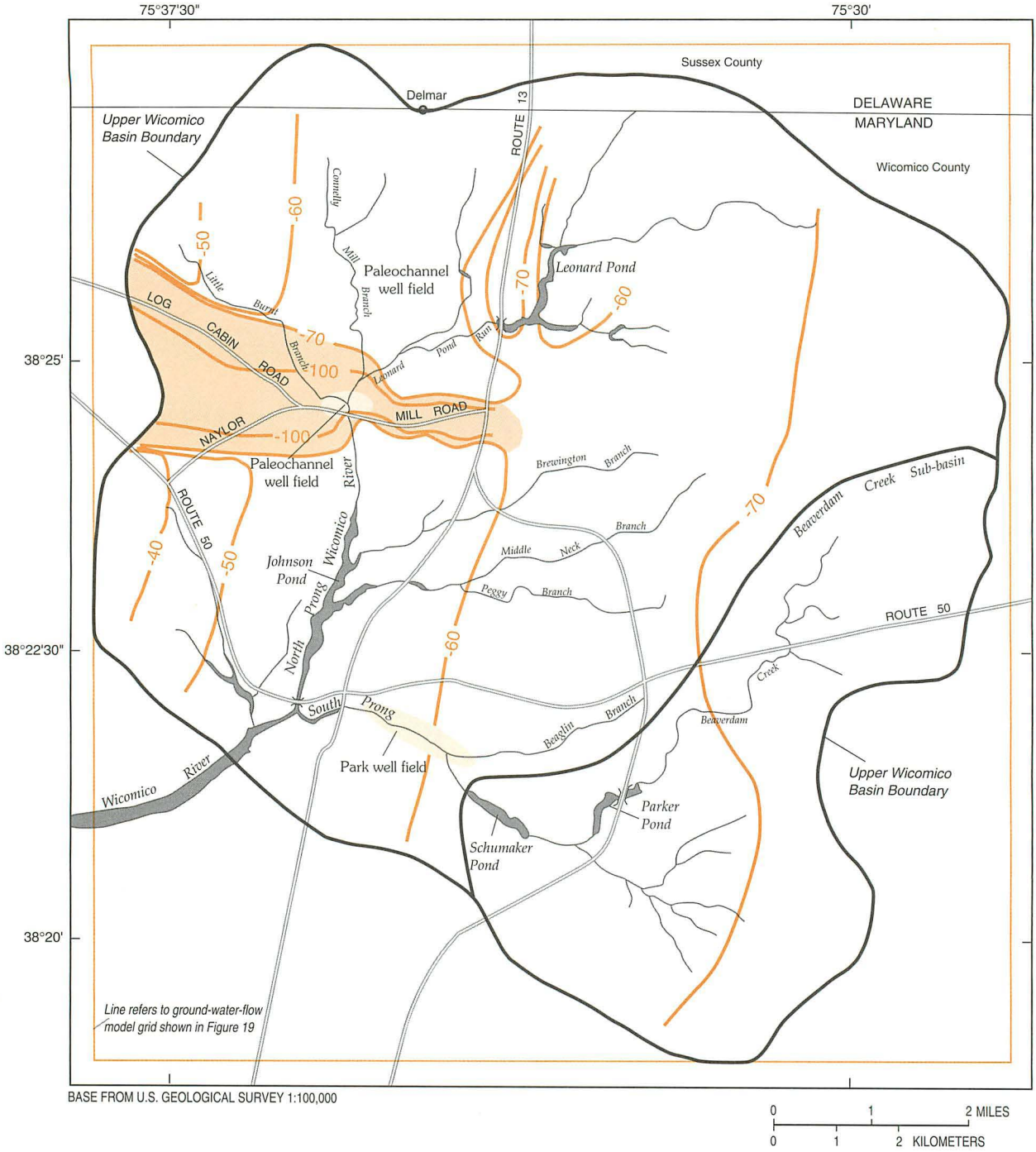
The presence of the Walston Silt and the Parsonsburg Sand on hydrogeologic cross sections A-A' and B-B' (figs. 3 and 4) are estimated from the Wicomico County geologic map (Owens and Denny, 1979a), because sufficient geophysical well-log data are not available.



EXPLANATION

-20 LINE OF EQUAL ALTITUDE, IN FEET RELATIVE TO SEA LEVEL. CONTOUR INTERVAL IS TEN FEET.

Figure 7. - Altitude of the bottom of the Beaverdam Sand.



EXPLANATION

-60 LINE OF EQUAL ALTITUDE, IN FEET BELOW SEA LEVEL. CONTOUR INTERVAL IS VARIABLE.

MAIN BRANCH OF THE SALISBURY PALEOCHANNEL. DEPTH TO BASE RANGES FROM 60 TO 200 FEET BELOW SEA LEVEL (Modified From Weigle, 1972).

Note: The structure of the paleochannel near Delmar is uncertain.

Figure 8. - Altitude of the bottom of the Salisbury aquifer.

WATER LEVELS AND DIRECTION OF GROUND-WATER FLOW

A map showing the altitude of the water table in the Salisbury aquifer was drawn using the median water levels recorded from selected wells between 1947 and 1994 (fig. 9). The map also includes single water-level measurements from different years. The water table is a subdued version of the topography. Water levels range in altitude from about 50 ft above sea level along the eastern edge of the basin to about sea level along the tidal portion of the Wicomico River. Generally, the ground water flows from the west, north, and east sides of the basin toward the tidal part of the Wicomico River. Water levels in the vicinity of the Paleochannel well field are about 15 to 20 ft above sea level, whereas in the Park well field the levels are less than 10 ft above sea level. Ground water flows from higher elevations near the edges of the basin in directions generally perpendicular to the contours of equal water-table altitude toward the major streams and toward the well fields where it is discharged. The water-level data is given in appendix C.

Long-term water-level records from four observation wells completed in the Salisbury aquifer were compared to the total estimated pumpage from the Salisbury aquifer. The observation wells are widely separated within the study area. Pumpage has increased gradually from approximately 2.6 Mgal/d for the period 1940-49 (Wheeler and Wilde, 1989) to approximately 8.7 Mgal/d in 1993. Generally, the hydrographs indicate that water levels fluctuate on a seasonal basis, but show no long-term effect from the increasing pumpage. High water levels typically occur in early spring and low levels occur in late fall. Water levels near the basin edges generally reflect recharge events more strongly than those near the major streams. Wells near streams, ponds, or pumping wells may show responses to local events, such as changes in stream stage, pond levels, or pumping rates.

Because the Salisbury aquifer is unconfined, has a high transmissivity, and is recharged rapidly from percolating precipitation, water levels respond to seasonal changes in recharge but show only small local responses to pumping (fig. 10). Aside from seasonal water-level changes and local adjustments to changes in pumping, the water-table surface has changed little over time.

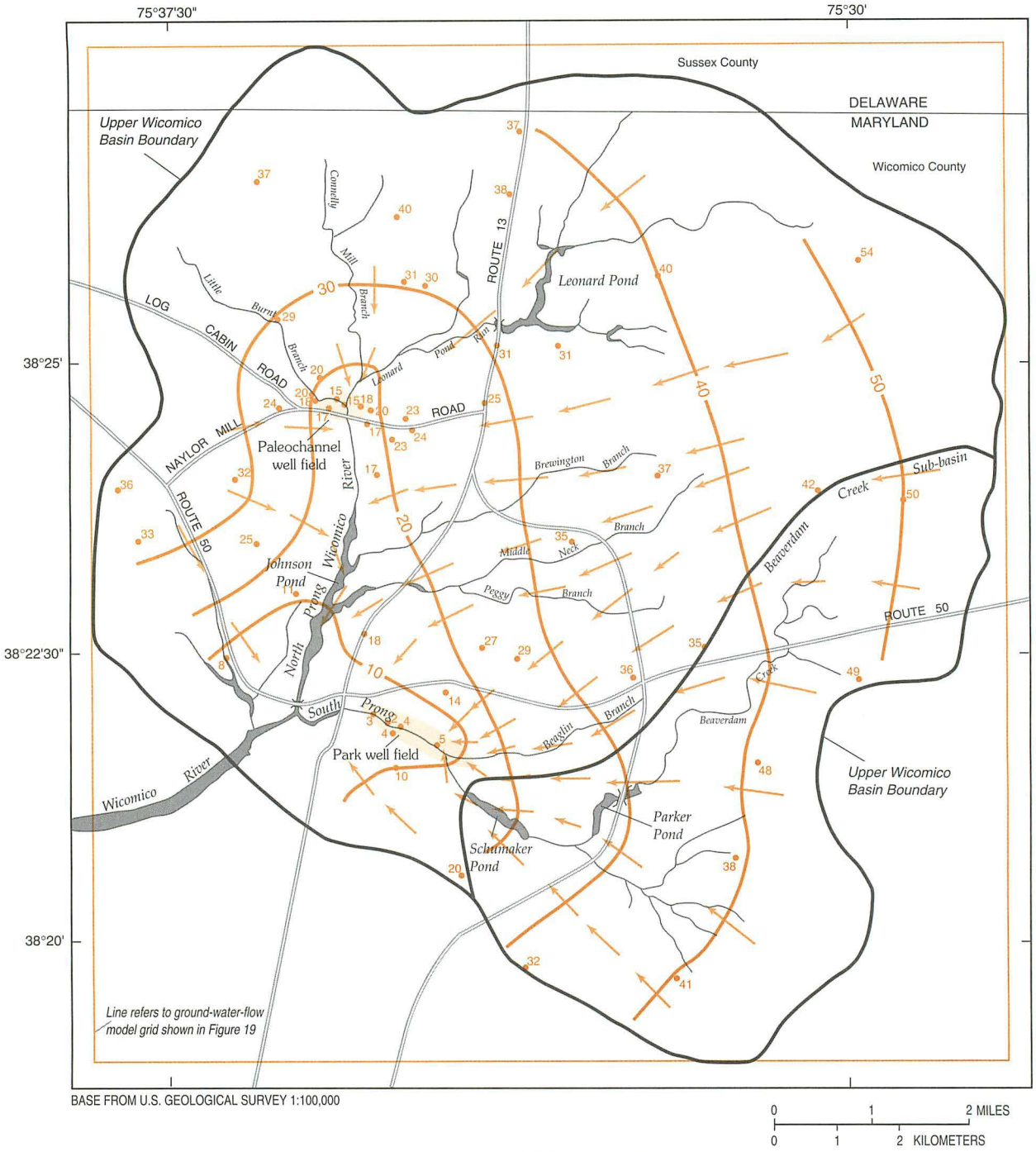
Water levels in well WI Cf 3, located at the Salisbury-Wicomico County Regional Airport, fluctuate as much as 6 ft between spring and fall. Pumping at the airport may also affect water levels in this well. Water levels in well WI Cf 147, located on the south side of Naylor Mill Road east of the Paleochannel well field, fluctuate seasonally as much as 5.5 ft.

Water levels in well WI Ce 204, located 45 ft from Paleochannel production well 1, fluctuated seasonally about 1 ft between 1968-75. In the fall of 1976 pumping from well 1 began, and as a result, water levels in well WI Ce 204 dropped about 6 ft below the pre-pumping low level of about 20 ft above sea level. Yearly water-level fluctuations in the well since 1976 have been as great as 8 ft. The trend in the data from 1976 to 1994 indicates no change between 1976-86 and a slight downward trend between 1986-94. The lower levels during the latter period are caused by an increase in average pumpage from 1.6 Mgal/d during 1976-86 to 2.9 Mgal/d during 1987-94.

Water levels in well WI Ce 13, located across the Park Pond from Park well 2a, show a less apparent seasonal correlation. Water levels are affected by pumpage from well 2a and by changes in the pond level. The aquifer is hydraulically interconnected to the pond, which typically is a ground-water discharge point. The pond is occasionally drained for maintenance. When this occurs, water levels drop in the aquifer.

Water levels measured between October 1993 and July 1995 in wells WI Ce 204, Ce 210, and Ce 213, located at distances of 45, 600, and 2,000 ft, respectively, from Paleochannel well 1 (WI Ce 200), are related to pumping at well 1 and precipitation recorded at the Salisbury-Wicomico County Regional Airport (fig. 11). Over most of the period of record the water-level trend in all three wells is relatively flat with the exception of an increase of about 2 ft during the late winter and spring of 1994. The rise in water levels during this period corresponds to an increase in precipitation during February and March 1994. A spike in precipitation (approximately 14 in.) occurred during July 1994 but was not reflected by a general rise in the ground-water levels, possibly because one or more storm events with fast rates of precipitation resulted in a greater percentage of runoff, and because soil moisture is taken up readily by plants during the growing season. Aside from the short-term water-level fluctuations resulting from the cycling of the pump in well 1, water levels do not respond to monthly changes in pumping. There is virtually no change in water levels from periods of low pumping to periods of high pumping (for example, between November 1994, when the pumping rate was 2.9 Mgal/d, to March 1995, when the pumping rate was 4.2 Mgal/d). Several downward spikes on the hydrograph for well WI Ce 213 and upward spikes on the hydrographs of wells WI Ce 210 and WI Ce 204 resulted from a momentary shift of pumping from production well 1 to production well 2. Production well 2 is located at a distance of about 200 feet from WI Ce 213.

Within the study area the direction of flow in the Manokin aquifer is generally towards the south (fig. 10). A cone-of-depression in the Manokin aquifer occurs in the area of Naylor Mill Road and the North Prong Wicomico River, where the Manokin and Salisbury aquifers are hydraulically interconnected by the paleochannel. The vertical head gradient is such that water flows from the Manokin aquifer to the Salisbury aquifer in this area of the paleochannel. This flow of water supplies the Salisbury aquifer with an additional source of recharge.



EXPLANATION

20 — WATER - TABLE CONTOUR, IN FEET ABOVE SEA LEVEL. CONTOUR INTERVAL IS TEN FEET.

48 MEDIAN WATER LEVEL FOR PERIOD OF RECORD, IN FEET ABOVE SEA LEVEL, (1947-94)(see appendix C).

GENERAL DIRECTION OF GROUND-WATER FLOW.

Figure 9. - Water-table altitude in the Salisbury aquifer.

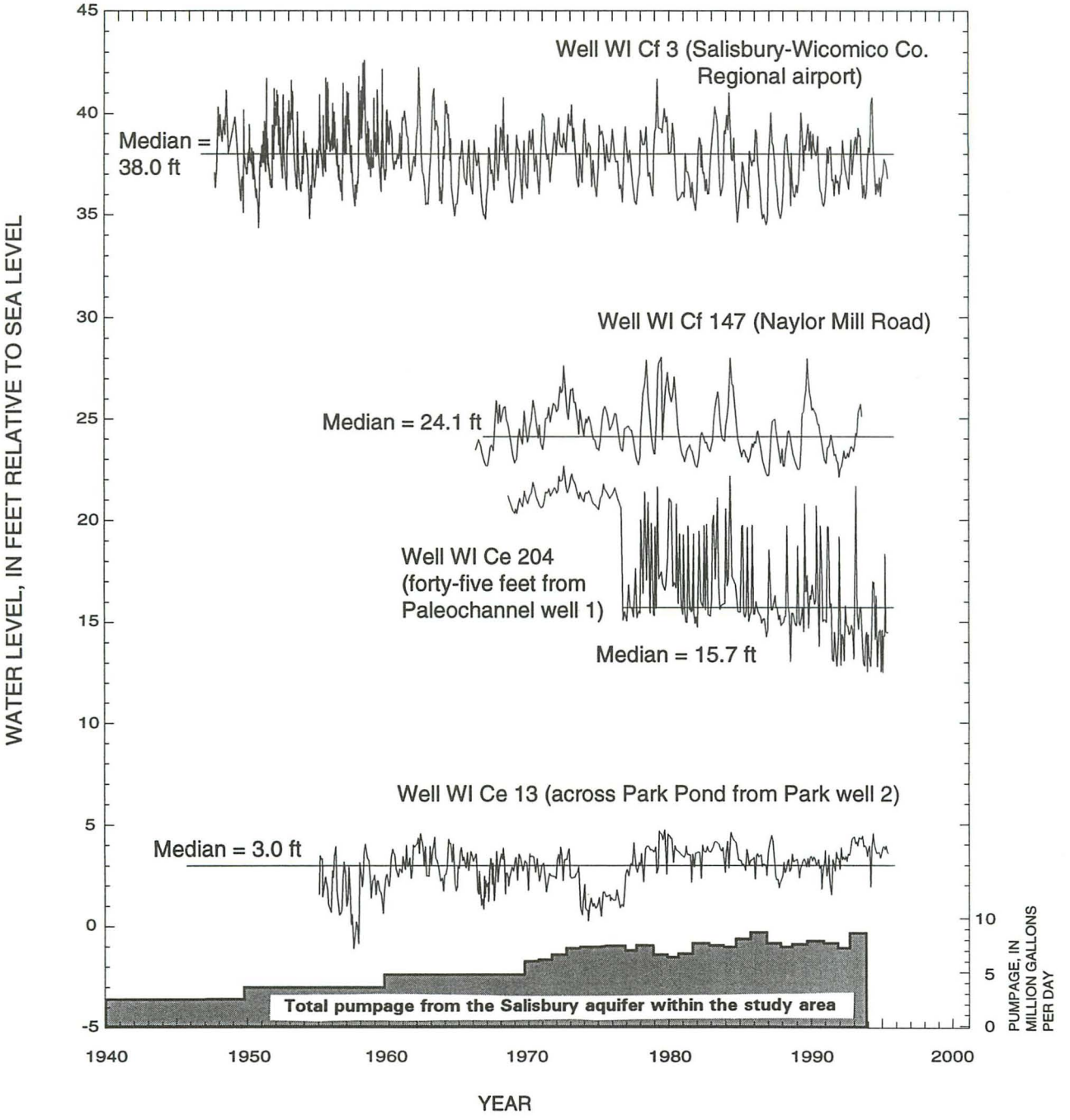


Figure 10. - Long-term water-level record from four observation wells screened in the Salisbury aquifer.

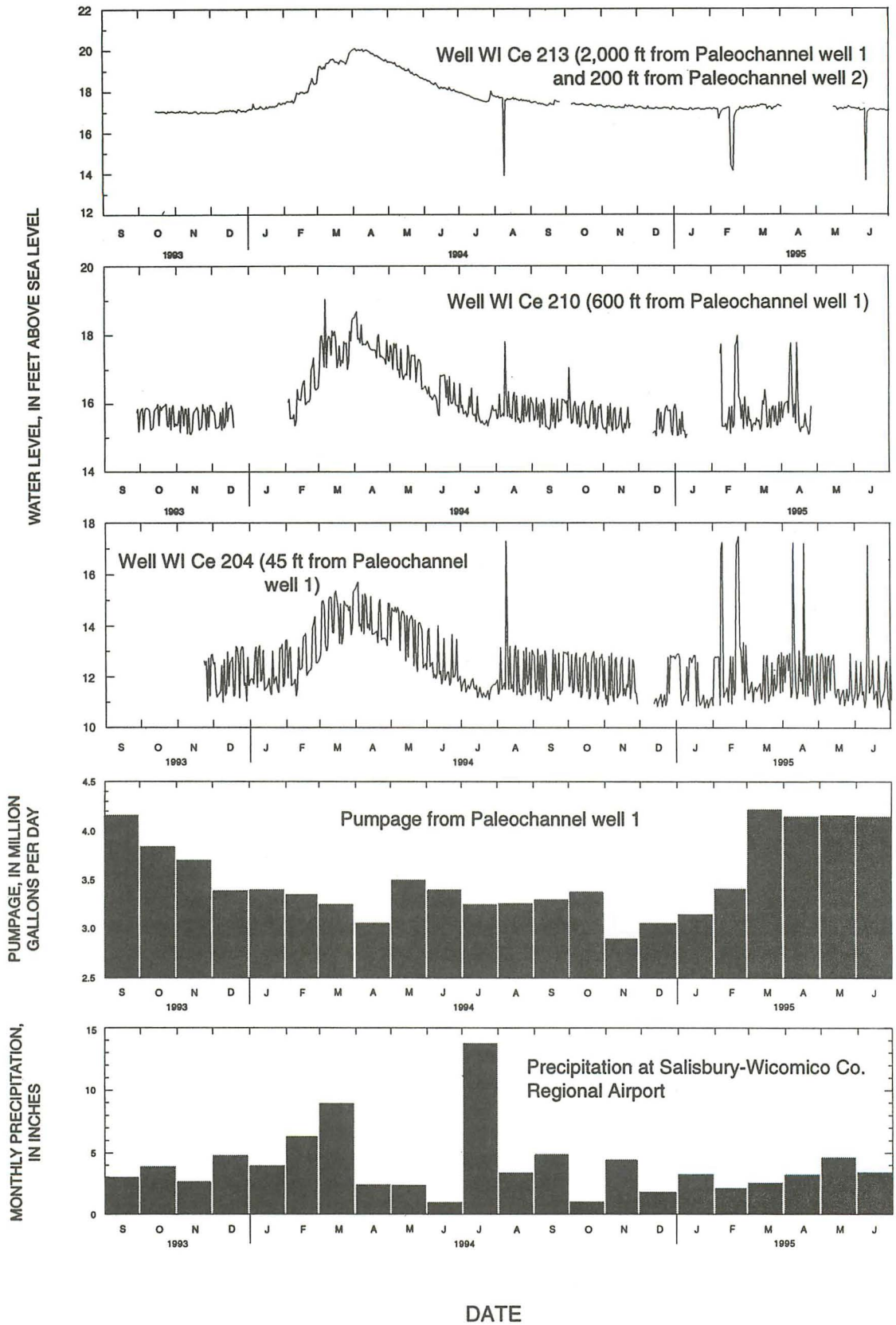
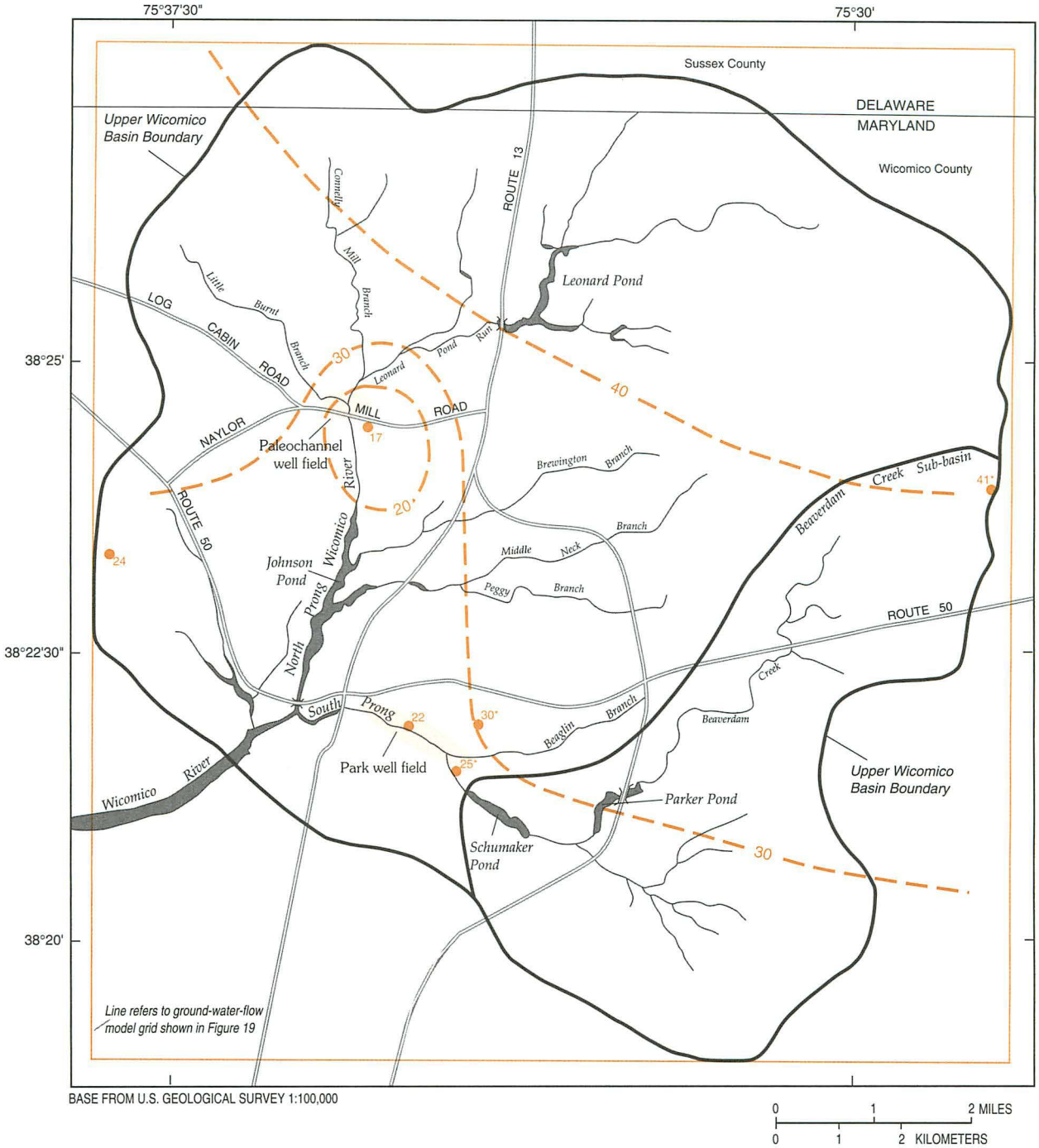


Figure 11. - Water-level fluctuations in wells WI Ce 204, 210, and 213 related to precipitation and pumpage from Paleochannel well 1.



EXPLANATION

30 — APPROXIMATE POTENTIOMETRIC SURFACE OF THE MANOKIN AQUIFER, IN FEET ABOVE SEA LEVEL. CONTOUR INTERVAL IS TEN FEET.

25 — WELL SCREENED IN THE MANOKIN AQUIFER. NUMBER IS THE WATER LEVEL IN FEET ABOVE SEA LEVEL. (* - indicates highest measurement of record (from Boggess and Heidel, 1968))

Figure 12. - Approximate potentiometric surface of the Manokin aquifer.

HYDRAULIC PROPERTIES

Horizontal Hydraulic Conductivity and Transmissivity

Data on the hydraulic properties of the aquifers and confining beds within the study area are limited mostly to the areas around the two city well fields. Little is known about the areal distribution of hydraulic properties of the aquifers elsewhere in the study area. Boggess and Heidel (1968) estimated the horizontal hydraulic conductivity of the Beaverdam Sand and Pensauken Formation ("red gravelly facies") to be about 230 feet per day (ft/d) and 350 ft/d, respectively. The estimates were based on sediment grain size contained in 233 core samples taken from 24 test holes completed in the project area. Assuming average thicknesses of 10 and 55 ft for the Beaverdam Sand and Pensauken Formation, respectively, at the Park well field, the estimated transmissivity (horizontal hydraulic conductivity times aquifer thickness) for the Salisbury aquifer using these horizontal hydraulic conductivities would be 22,000 ft^2/d.

Aquifer tests in the Salisbury aquifer conducted near Park well field (Rasmussen and Slaughter, 1955) indicated transmissivities ranging from about 8,200 to 57,000 ft^2/d and an average of 13,000 ft^2/d.

Slug tests in the upper part of the Salisbury aquifer (Beaverdam Sand) at the Wicomico County Newland Park landfill (fig. 1) indicates that the horizontal hydraulic conductivity may be lower in that area. Results of the tests in four wells screened at depths ranging from 16 to 27 ft below land surface show horizontal hydraulic conductivities ranging from 11 to 196 ft/d (GeoSyntec Consultants, 1992). The average value was 76 ft/d.

The paleochannel sediments, corresponding to Zone B of the Pensauken Formation (tab. 1), have in general the highest transmissivity values of all the layers considered in this report. A test of the water-bearing properties of these sediments was conducted by Mack and Thomas (1972) in 1967. During their study, well WI Ce 200 (Paleochannel well 1) was pumped at 4,000 gallons per minute (gal/min) for 30 days. A transmissivity of 53,500 ft^2/d was calculated using the Theis-type curve method on drawdown and recovery data in the pumping well and two observation wells. This value was thought to represent the maximum for the aquifer since both pumping and observation wells were located in the thickest section of sand and gravel. This value is probably high because some water undoubtedly flowed from sands above and below the screened interval (Zone B) (tab. 1).

No aquifer test data are available for the lower part of the paleochannel (Zone C) (tab. 1). However, Mack and Thomas (1972) indicate that the lithology consists of fine to coarse sand capable of yielding moderate amounts of water. This deposit may consist of reworked Manokin sediments, in which case the hydraulic conductivity may be similar to that of the Manokin aquifer.

The Manokin aquifer is moderately productive within the study area. Boggess and Heidel (1968) calculated a transmissivity value of 5,350 ft^2/d at the Mardel By-Products Plant (well WI Ce 148; pl. 1) west of Salisbury. The approximate horizontal hydraulic conductivity calculated using this transmissivity value and an approximate aquifer thickness of 100 ft is 54 ft/d. The transmissivity of the Manokin decreases southward. Analysis of an aquifer test at the Nutters Crossing Golf Course (wells WI Df 62 and 63; pl. 1), located just south of the study area, shows a transmissivity between 1,700 and 2,400 ft^2/d. In that test, water levels in well WI Df 63 were observed while well WI Df 62, 12 ft away, was pumped. Both wells are screened between approximately 110 and 200 ft below sea level. Transmissivity was determined by the Jacob straight-line method using the procedure described in Driscoll (1986).

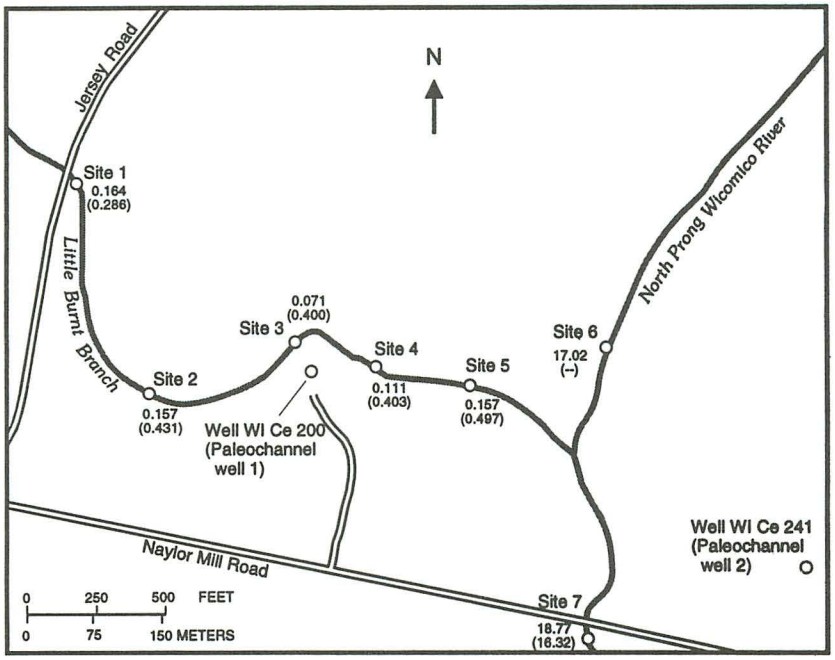
Vertical Hydraulic Conductivity of the Lower Confining Bed

Vertical hydraulic conductivity of the confining bed separating the Salisbury aquifer from the Manokin aquifer was taken from field and laboratory analyses of core material near the Park well field (Wolff, 1970). Wolff calculated hydraulic diffusivity (K/S, where K = hydraulic conductivity and S = specific storage) in the field (aquifer test) and in the laboratory (consolidation and constant flow tests). The tests resulted in similar values. Vertical hydraulic conductivity determined using the constant flow and consolidation methods resulted in values ranging from 2.8×10^{-5} to 6.2×10^{-3} ft/d and 1.1×10^{-4} to 1.8×10^{-3} ft/d, respectively. The median values were 2.7×10^{-4} and 5.4×10^{-4} ft/d, respectively.

GROUND-WATER/SURFACE-WATER INTERACTIONS

Stream flows were measured during baseflow periods to determine rates of ground-water seepage. Baseflow is the ground-water contribution to stream flow. Baseflow seepage runs were made at five sites along Little Burnt Branch near Paleochannel well 1 (WI Ce 200) and at two sites along the North Prong Wicomico River (fig. 13) on December 3, 1993 and October 19, 1994. The purpose of the tests was to determine whether current pumping rates were diverting baseflow from the streams. Both tests were conducted after a minimum of 5 days of zero precipitation. This was necessary to separate baseflow from runoff in total streamflow.

The 1993 measurements indicate that pumping from Paleochannel well 1 (WI Ce 200) lowered baseflow by 0.1 cubic feet per second (ft^3/s), or about 45 gallons per minute (gpm), between the upstream site at Jersey Road (site 1) and the site near the production well (site 3) (fig. 13). This accounts for a 60-percent reduction in baseflow between



EXPLANATION

0.164 (0.286) — BASEFLOW, IN CUBIC FEET PER SECOND. TOP NUMBER IS FROM DECEMBER 3, 1993 AND BOTTOM NUMBER IS FROM OCTOBER 19, 1994.

Site 1 — SITE AND NUMBER

Figure 13. - Baseflow measured along Little Burnt Branch near the Paleochannel well field.

these two sites. Baseflow increased to 0.157 ft^3/s at the downstream site (site 5). The 1994 measurements show an increase in baseflow from site 1 to the next downstream site (site 2), then a slight decrease as the stream passes the production well at site 3. Baseflow then gradually increases again downstream. The Paleochannel well field produced an average of 3.3 and 3.4 Mgal/d during December 1993 and October 1994, respectively. Some of the water was pumped from well 2 (WI Ce 241) located on the east side of the North Prong Wicomico River. Baseflow along the North Prong Wicomico River, measured upstream of Little Burnt Branch (site 6) and just south of Naylor Mill Road (site 7), is two orders-of-magnitude greater than that measured in Little Burnt Branch (fig. 13).

Hydraulic potentiomanometer readings were made concurrently with the 1994 baseflow measurements to determine the direction of seepage by measuring the head difference between the stream and the ground water immediately below the stream. The potentiomanometer consisted of a stainless-steel potentiometer with a sliding outer sleeve that covered a fine mesh screen located at the bottom of the tube. After the tube was inserted into the streambed, the sleeve was lifted to expose the screen. The manometer consisted of flexible plastic tubing connected to the potentiometer at one end and open to the stream at the other end. A tube and valve at the top of the manometer ran to a peristaltic pump. The pump was used to fill the potentiometer side of the manometer. This also allowed for collection of ground-water samples. A detailed discussion of the design and use of this type of hydraulic potentiomanometer is given in Winter and others (1988).

Aquifer heads were greater than stream heads at sites 1, 2, 4, and 5 (fig. 14). Head differences between the aquifer and stream ranged from 0.16 to 0.37 ft at these sites. Head in the aquifer was lower than head in the stream by 0.29 ft at site 3, which is adjacent to production well 1. These head

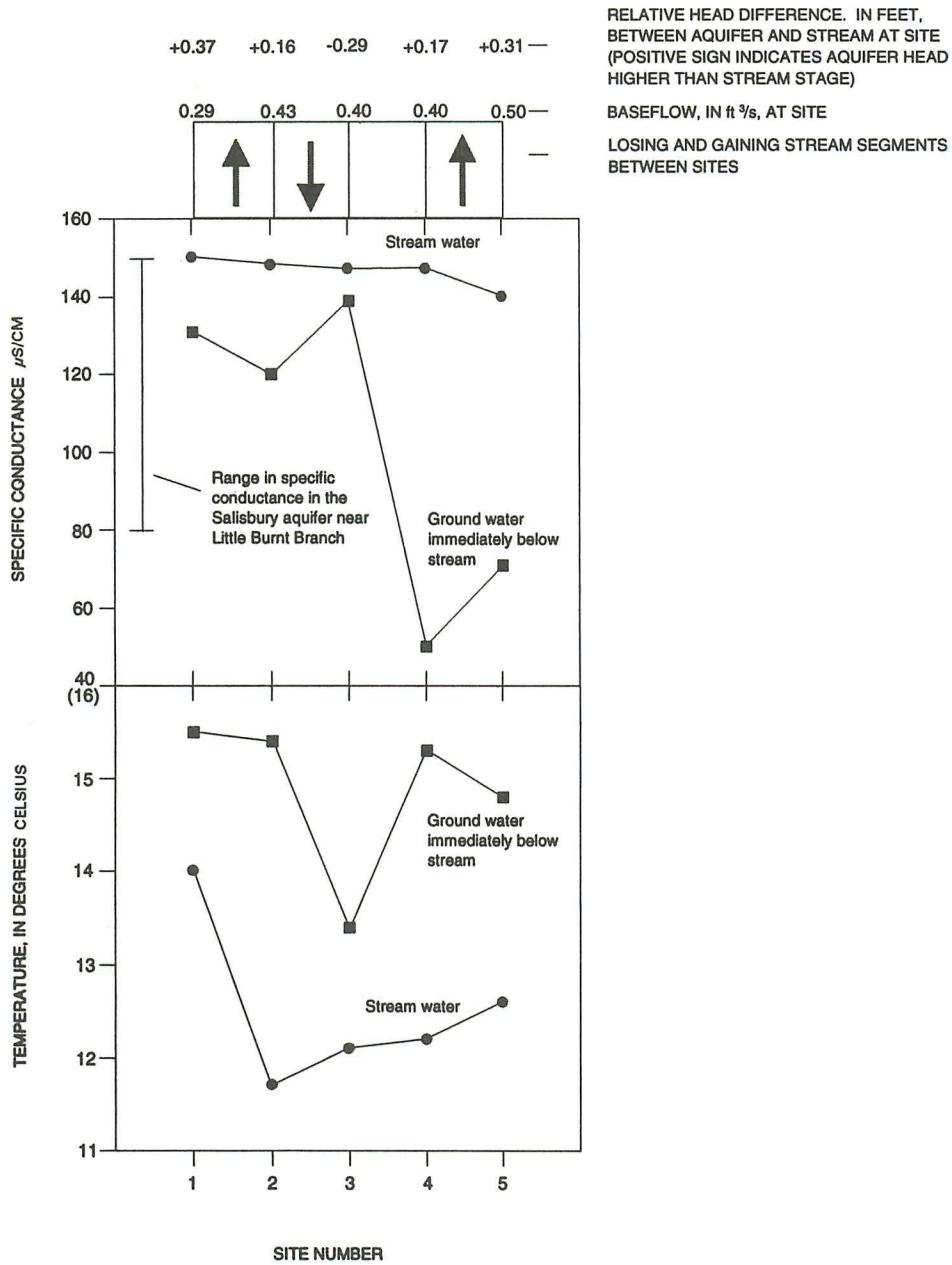


Figure 14. - Relative head distribution, specific conductance, and temperature measured in Little Burnt Branch and the underlying stream bed during the October 19,1994 seepage run.

relations indicate seepage out of the stream as a result of pumping at production well 1.

Temperature and specific conductance were measured in both the stream and in ground water pumped from the potentiometer to further understand the ground-water/surface-water interactions (fig. 13). Stream-water temperature remained fairly consistent (11.7°C to 12.6°C) with the exception of site 1, where the temperature was 14.0°C. The higher temperature at site 1 could be caused by a greater influx of warmer aquifer water. Aquifer water temperature was similar (14.8°C to 15.5°C) at all sites except site 3, which is adjacent to the production well. Aquifer water temperature at that site was 13.4°C, which suggests that there is infiltration of cooler stream water at that location. Specific conductance of the stream was fairly similar at all sites (140 to 150 microsiemens per centimeter at 25°C [µS/cm]). Specific conductance in the aquifer was nearly identical at the first two upstream sites at about 125 µS/cm, then increased at site 3 (near production well 1) to stream water levels. Downstream of site 3, the specific conductance decreased to 50 and 71 µS/cm at sites 4 and 5, respectively. The increase at site 3 could be a result of mixing of stream and aquifer water. The cause of the decrease in specific conductance at sites 4 and 5 is unclear; but it may represent the natural variation in aquifer water conductance. Specific conductance measured in water from the Salisbury aquifer near Little Burnt Branch ranged between 116 µS/cm in well WI Ce 196 (fig. 2), screened between 6 and 7 ft below land surface, and 80 µS/cm in Paleochannel well 1 (WI Ce 200), screened between 80 and 160 ft below land surface. Ground-water-flow paths from different depths converge at discharge areas such as streams. It is possible that a deeper flow path was tapped at sites 4 and 5.

The hydraulic connection between the Park Pond and the Salisbury aquifer was tested by Boggess and Heidel (1968) by pumping a production well adjacent to the pond (well 2--replaced by well 2a) while observing water levels in surrounding wells. Results from the test indicate that the effect of pumping (drawdown) was observed in observation wells on the opposite side of the pond. This effect could be attributed to a localized, semi-confined condition caused by the presence of thin clay layers within the aquifer, or by low-permeability pond sediments, which prevent pond water from recharging the aquifer. Split-spoon core samples of Park Pond bottom sediment (fig. 15), however, show predominantly fine- to coarse-grained sand (app. B). Therefore, anisotropic aquifer sediments are more likely to have allowed for the transfer of pressure decline across the pond. In another test by Boggess and Heidel (1968), discharge measurements made along the upper part of the South Prong Wicomico River between Park well 7b (WI Ce 247) and Schumaker Pond indicated a decrease in baseflow along the reach that passed pumping wells in the well field. This indicates the existence of a direct connection between the stream and aquifer.

In summary, surface water is an important factor governing ground-water flow near pumping wells in both the Paleochannel and Park well fields. Wells pumping near the surface-water bodies can capture baseflow, or if the pumpage is great enough, can induce recharge from the surface water. Anisotropy within the aquifer sediments may, however, limit the amount of flow between aquifer and surface water.

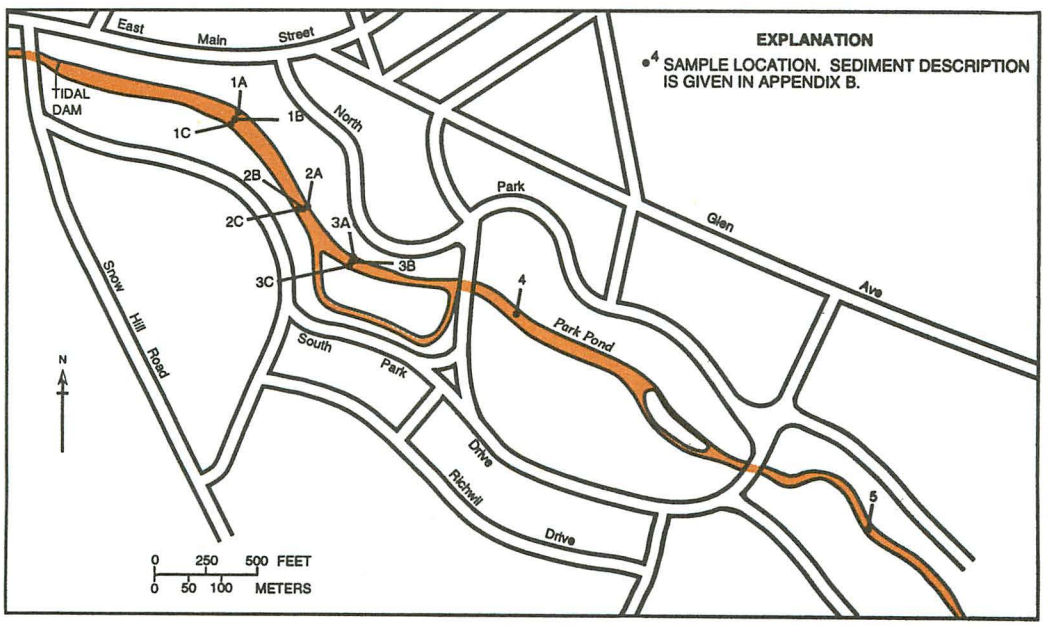


Figure 15. - Location of bottom-sediment cores in Park pond.

RECHARGE AND BASEFLOW

Ground-water recharge in this report refers to the net volume of water entering the Salisbury aquifer after evapotranspiration is subtracted. This volume of water is sometimes referred to as effective recharge. Recharge to the Salisbury aquifer within the Upper Wicomico River Basin used in this report was estimated to be 13.4 inches per year (in/yr) based on baseflow separation from stream flows recorded in the Beaverdam Creek sub-basin (fig. 1). No other long-term stream flow record exists within the basin. Equating baseflow to effective recharge is limited in that it assumes that no water is withdrawn by wells and that all water flowing to the stream received recharge from within the stream basin and not from deeper aquifers or areas outside the basin. Also, using a recharge value representative of a subbasin for the larger basin assumes that the hydrologic characteristics of the subbasin are similar to the entire basin--a condition which may not be met. Streamflow in Beaverdam Creek used for estimating recharge was measured at the outlet of Schumaker Pond by the U.S. Geological Survey between 1931-76. Baseflow was separated from the streamflow hydrograph by computer (White and Sloto, 1990) using three techniques--fixed-interval, sliding-interval, and local-minimum (Pettyjohn and Henning, 1979). The median value of baseflow in the Beaverdam Creek Basin for the entire period of record calculated by the fixed-interval and sliding-interval techniques were 19.31 ft^3/s and 19.29 ft^3/s, respectively. Median baseflow in the basin computed by the local-minimum technique was 17.67 ft^3/s. This technique generally gives a more conservative (lower) estimate of baseflow (White and Sloto, 1990). An average of 19.30 ft^3/s from the first two methods was chosen to represent baseflow in Beaverdam Creek. This value is equivalent to 13.4 in/yr of recharge over the basin area of 19.5 mi^2.

Estimates of baseflow in five basins in the central part of the Delmarva Peninsula with hydrogeologic characteristics similar to the Upper Wicomico River Basin were made for comparison. The median recharge for the five selected basins ranged from 9.5 to 14.2 in/yr (tab. 2). Baseflow was separated from stream-flow hydrographs using the techniques mentioned above. The recharge rates (equivalent to baseflow) determined at these sites are all within 4 in/yr of the rate estimated for the Upper Wicomico River Basin.

A comparative recharge value is found in a comprehensive study of the hydrologic budget of the Beaverdam Creek Basin by Rasmussen and Andreasen (1959). In that study recharge (effective) was calculated at 25.9 percent of precipitation for the period 1950-52. The average annual precipitation measured at the Salisbury-Wicomico County Regional Airport (fig. 1) between 1948-93 is 44 in. Based on this value, the recharge rate (25.9 percent of precipitation) is 11.4 in/yr.

Another source of recharge to the Salisbury aquifer is from the Manokin aquifer within the paleochannel area. Water from the Manokin aquifer discharges through the paleochannel to the Salisbury aquifer, contributing to baseflow in Little Burnt Branch (Boggess and Heidel, 1968). The Manokin receives recharge west of the study area where it is hydraulically interconnected to the Salisbury aquifer (Boggess and Heidel, 1968).

Table 2. - Baseflow for selected streams on the central Delmarva Peninsula outside the study area and for Beaverdam Creek

[ft^3/s, cubic feet per second; in/yr, inch per year]

Site name	U.S. Geological Survey site number	Median baseflow (ft^3/s)	Median recharge to basin (in/yr)
Faulkner Branch at Federalsburg, Md.	01489000	6.3	12.0
Nanticoke River near Bridgeville, Del.	01487000	66.9	12.0
Nassawango Creek near Snow Hill, Md.	01485500	32.6	9.7
Pocomoke River near Willards, Md.	01485000	42.4	9.5
Stockley Branch at Stockley Branch, Del.	01484500	5.43	14.2
Beaverdam Creek at Schumaker Pond, Md.	01486500	19.3	13.4

GROUND-WATER USE

Total ground-water use from the Salisbury aquifer within the study area was approximately 8.7 Mgal/d in 1993 (Judith Wheeler, U.S. Geological Survey, oral commun., 1995). Public, commercial, industrial, and irrigation users withdrawing over 10,000 gallons per day (gal/d) are required to report monthly pumpage totals, but irrigation use was estimated from the average appropriated amounts, because reported pumpage figures were not available for 1993. The City of Salisbury pumped approximately 3.5 and 1.7 Mgal/d from the Paleochannel and Park well fields, respectively. The total amount pumped by the city accounted for approximately 60 percent (5.2 Mgal/d) of the total water withdrawn within the basin. Commercial and industrial use accounted for 21 percent (1.8 Mgal/d), irrigation use accounted for 11 percent (1.0 Mgal/d), and other public and private (other than domestic) supplies accounted for 0.4 percent (or 0.04 Mgal/d). Domestic supplies accounted for approximately 0.7 Mgal/d. A list of ground-water appropriations greater than 10,000 gal/d from the Salisbury aquifer in the Upper Wicomico River Basin is given in appendix D.

Salisbury's Paleochannel well field pumps water from two large capacity wells screened in the Salisbury aquifer. Each well is equipped to pump between 2,000 and 2,200 gallons per minute (Earl Wells, City of Salisbury Water Plant, oral commun., 1995). The older of the two wells--well 1 (well WI Ce 200)--is used more frequently than well 2 (WI Ce 241), because water pumped from well 2 contains higher concentrations of dissolved iron and, therefore, requires more treatment.

Salisbury's Park well field pumps water from nine production wells screened in the Salisbury aquifer (fig. 2). The wells are completed to a depth of approximately 60 ft below land surface. The current pumping capacities of the wells range from 200 to 1,300 gal/min and average around 700 gal/min (William Dodson, City of Salisbury Water Plant, oral commun., 1995).

Individual residential ground-water users in a number of communities located on the outskirts of the Salisbury public-supply service area pump water from the Salisbury aquifer. These communities account for most of the privately-owned wells within the study area. The amount pumped for domestic supply was estimated at 0.7 Mgal/d by counting houses on the 1992 Delmar and 1982 Salisbury quadrangles and multiplying that figure by an average daily use per household of 300 gal/d. The majority of residential wells pump from the Salisbury aquifer.

The Manokin aquifer supplies water to a relatively small number of residential, irrigation, and commercial wells within the basin. There is no appropriated use over 10,000 gal/d for this aquifer located within the basin.

GROUND-WATER AGE

Knowledge of the age of ground water is useful in understanding the ground-water-flow system. In particular, it can be used to evaluate and improve the performance of ground-water-flow models by providing an independent estimate of ground-water velocities (Riley and others, 1994).

The Salisbury aquifer is a shallow, permeable unit that is probably recharged rapidly. Several methods exist for determining the age of recently recharged ground water. One such method is the use of environmental tracers (Plummer and others, 1993). Many different chemicals have been emitted into the atmosphere by man since the beginning of the industrial era. Some of these chemicals become partitioned in the hydrologic cycle and enter the ground-water system during recharge events. If the chemical is relatively stable in, and can flow through, the hydrologic system, then the chemical can be used as a ground-water tracer. Generally, the concentration of the chemical in ground water can be matched to the date when those concentrations occurred in the atmosphere. Two types of environmental tracers were used in this study--chlorofluorocarbons (CFC-11, CFC-12, and CFC-113) and tritium (^3H). These tracers were chosen because they make it possible to identify waters recharged within about the last 50 years (Plummer and others, 1993). Also, using both tracers allows for comparison of results.

Sampling of discrete zones within the aquifer produces the best results, because fewer flow paths are intercepted, which reduces the amount of mixing of different age ground water. This condition was met in only a few wells sampled during this study. Most wells sampled had relatively long sections of well screens that produce water of different ages.

Age Determined by Chlorofluorocarbon Analysis

Chlorofluorocarbons (Freons), specifically dichlorodifluoromethane (CFC-12) and trichlorofluoromethane (CFC-11) are stable volatile organic compounds that were first produced in the 1930's and 1940's, respectively (Plummer and others, 1993). Trichlorotrifluoroethane (CFC-113) is another chlorofluorocarbon compound of commercial significance. CFC-11 and CFC-12 make up 77 percent of the total CFC global market, with the remainder composed of CFC-113 and other less commercially important CFC compounds (Busenberg and Plummer, 1992). CFC's are used as refrigerants, cleaning agents, aerosol propellants, and blowing agents in the production of plastics (Plummer and others, 1993). Eventually, CFC's are released to the atmosphere, where their concentrations have been gradually increasing since the 1930's (fig. 16).

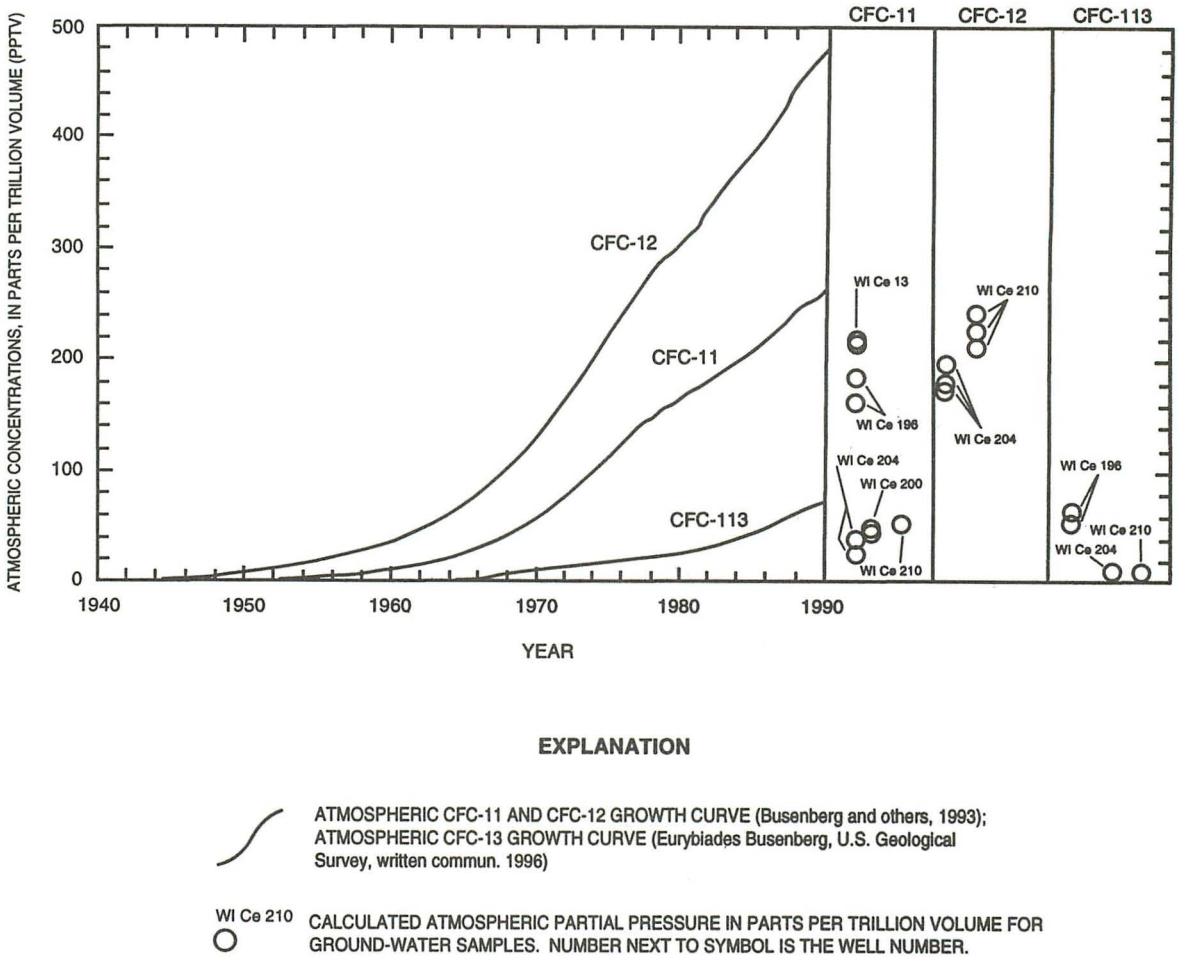


Figure 16. - Comparison of chlorofluorocarbon (CFC) concentrations in the Salisbury aquifer with the atmospheric CFC growth curve.

The concentrations of CFC's in solution are linearly related to the partial pressure of the CFC's as expressed through Henry's law:

$$C = k\,P,$$

where C is the concentration of the CFC compound in water, k is the proportionality constant--called Henry's law constant--for the compound, and P is the partial pressure of the CFC compound in a gas phase (air) in equilibrium with the water. The solubility of CFC, as given in k, is dependent on pressure and recharge temperature. The partial pressures calculated from Henry's law are compared to atmospheric growth curves for CFC's (Dunkle and others, 1993) to determine the year in which the compound entered the ground-water system. Gas exchange between the saturated and unsaturated zones occurs under atmospheric pressure. The recharge temperature--or the temperature at the base of the unsaturated zone--varies with depth to the water table. In aquifers in which the unsaturated zone is relatively thin, such as the Salisbury aquifer, the recharge temperature also varies seasonally (Plummer and others, 1993). A recharge temperature of 12.0°C, similar to those determined on the Delmarva Peninsula by Dunkle and others (1993), was used in this study. A maximum age error of several years results from uncertainties of several degrees Celsius in recharge temperature when using CFC-11 and CFC-12 in the young waters of the study area (1980-90) (Plummer and others, 1993).

Factors that can increase or decrease the concentrations of chlorofluorocarbons in ground water include, (1) contamination from local CFC sources and from the atmosphere during sampling, (2) mixing in well casings, (3) hydrodynamic dispersion, (4) microbial degradation, (5) soil sorp-

tion and desorption, and (6) entrainment of excess air during recharge (Dunkle and others, 1993). Factors (1), (2), and (3) are likely to have the greatest impact on concentrations in the Salisbury aquifer.

Water samples from the Salisbury aquifer were collected from seven wells (between November 9 and December 1, 1993) and analyzed for CFC concentrations using methods of Busenberg and Plummer (1992) (tab. 3). Wells sampled for CFC's in the Paleochannel well field included WI Ce 196, WI Ce 200 (production well 1), WI Ce 204, WI Ce 210, and WI Ce 242 (fig. 2). Water from WI Ce 196--located between Little Burnt Branch and production well 1--was recharged around 1983. The water was thought to be slightly contaminated with CFC-12. The term "contamination" is defined here as water with CFC concentrations greater than the recharge air-water equilibrium concentration (at 12.0°C) with the 1993 atmosphere--approximately 328 picograms per kilogram (pg/kg) and 723.5 pg/kg for CFC-12 and CFC-11, respectively. The additional amount of CFC's in solution may be the effect of higher CFC concentrations often present in urban air and runoff. Assuming too high a recharge temperature could also explain, in part, the higher CFC concentrations. Water produced from well WI Ce 196 contained 13 milligrams per liter (mg/L) of dissolved chloride. Levels over 10 mg/L in shallow ground water generally result from surficial contamination which suggests relatively short ground-water flow paths.

Water from well WI Ce 200 was thought to have been recharged about 25 years ago (1968) based on CFC-11 and CFC-113 concentrations; however, water from this well was highly contaminated with CFC-12 from an unknown source. The recharge date for water produced from well WI Ce 204 is about 1970 based on CFC-11, CFC-12, and CFC-113 concentrations. Well WI Ce 210 was also recharged in the early 1970's. Water produced from this well contained 4.1 mg/L of dissolved nitrate plus nitrite (as nitrogen). Nitrate plus nitrite concentrations greater than 3.0 mg/L suggest contamination from anthropogenic sources (Bachman, 1984); the presence of nitrogen above 3.0 mg/L may indicate relatively short ground-water flow paths. Water from well WI Ce 242, adjacent to Production well 2, was highly contaminated with CFC-12 from an unknown source, but the water contained very small amounts of CFC-11 and CFC-113, suggesting that it may be more than 30 years old (recharged prior to 1963).

In general, older water occurs deeper in the aquifer at the well field. This is probably because deep ground-water-flow paths have long travel times in the Salisbury aquifer and could be mixed with older water from the Manokin aquifer beneath.

Wells sampled in the Park well field included WI Ce 13 and WI Ce 244 (production well 2a). Well WI Ce 13 was highly contaminated with CFC-12 and contained other constituents that masked CFC-113. A recharge date of 1986 was assigned based on CFC-11 concentrations. Water from well WI Ce 244 was highly contaminated with CFC-11 and CFC-12 from an unknown source and contained another organic compound, which interfered with the CFC-113 analysis; therefore, a recharge date could not be assigned.

Age Determined by Tritium Analysis

Tritium (^3H) is a radioactive isotope of hydrogen. It is formed naturally from cosmic-ray spallation in the upper atmosphere (Robertson and Cherry, 1989). Background concentrations of tritium in the atmosphere range from 2 to 8 tritium units (TU) (Plummer and others, 1993). Starting in 1953, large amounts of tritium were introduced into the atmosphere by atmospheric testing of nuclear weapons. The amount released to the atmosphere far exceeded background concentrations. Atmospheric tritium concentrations peaked during the large-scale weapons testing in 1962-63 and has declined since then (fig. 17). High concentrations of tritium in the atmosphere in concert with its chemical stability and the fact that it substitutes for hydrogen (^1H) in the water molecule--thus allowing it to travel at the same rate as water in the ground-water system--makes tritium an excellent tracer for ground water recharged within the past 40 years. The half-life of tritium is 12.43 years (Plummer and others, 1993). To compare tritium concentrations in ground water to previous atmospheric levels, the atmospheric concentrations must be corrected for radioactive decay using the equation:

$$\ln A = \ln A_o - \frac{t}{t_{1/2}(1.44)},$$

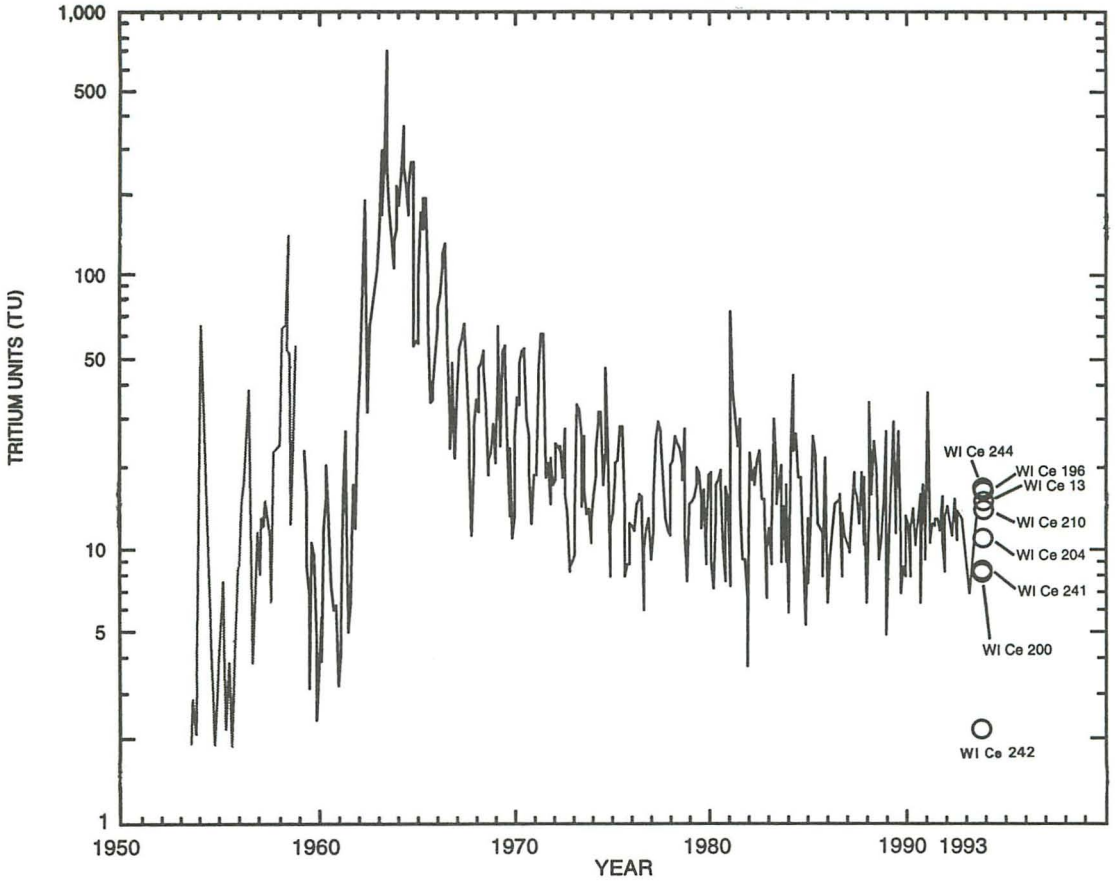
where A is the atmospheric tritium concentrations corrected for radioactive decay (in TU), A_o is the concentration at the time of isolation from the atmosphere (in TU), t is the time since isolation (in years), and $t_{1/2}$ is the half-life of tritium (12.43 years) (Coplen, 1993); one tritium unit equals 3.2 picoCuries per liter (pCi/L). When estimating recharge dates using tritium, it is assumed that the date of recharge coincides with the date that the precipitation was isolated from the atmosphere--when it first entered the ground.

Water samples were collected and analyzed for tritium concentrations in eight wells in the Salisbury aquifer--six in the Paleochannel well field and two in the Park well field--between November 9, and December 1, 1993. The wells sampled in the Paleochannel well field include WI Ce 196, WI Ce 200 (production well 1), WI Ce 204, WI Ce 210, WI Ce 241 (production well 2), and WI Ce 242 (fig. 2). The wells sampled in the Park well field include WI Ce 244 (production well 2a) and WI Ce 13 (fig. 2). Estimated recharge dates were assigned by comparing tritium concentrations in the samples to the decay-corrected tritium concentrations in precipitation collected at Washington, D.C. and Ottawa,

Table 3. - Chlorofluorocarbon data from water samples collected in wells screened in the Salisbury aquifer

[Recharge temperature is 12 C; pg/kg, picograms per kilogram; pptv, parts per trillion volume; CFC, chlorofluorocarbon; ERR, interference from other chemical constituents; Contam., contaminated with CFC above 1993 atmospheric concentrations]

Well number	Date	Concentrations in water (pg/kg)			Calculated partial pressures (pptv)			CFC-model recharge date			Comments (Eurybiades Busenberg, U.S. Geological Survey, written commun., 1994)
		CFC-11	CFC-12	CFC-113	CFC-11	CFC-12	CFC-113	CFC-11	CFC-12	CFC-113	
WI Ce 13	12/1/93	578.3	866.1	ERR	221.8	1,437.4	ERR	1985.5	Contam.	ERR	Highly contaminated with CFC-12. Interference-
		585.3	851.5	ERR	216.8	1,413.1	ERR	1985.0	Contam.	ERR	cannot use CFC-113. Recharge date 1985?
		573.3	858.5	ERR	219.8	1,424.7	ERR	1985.5	Contam.	ERR	
WI Ce 196	11/23/93	476.5	338.9	56.1	182.8	562.4	52.4	1981.5		1986.0	May be slightly contaminated with CFC-12.
		478.4	335.8	56.7	183.5	557.3	52.9	1981.5		1986.0	Water is about 10 years old.
		479.9	323.5	57.9	184.1	536.9	54.1	1981.5		1986.0	
		1,013.7	858.4	69.4	388.8	1,424.5	64.8	Contam.	Contam.	1988.0	
WI Ce 200	11/23/93	754.7	966.1	50.2	289.5	1,603.3	46.9	Modern	Contam.	1985.0	Well not purged for sample number 2. Highly
		125.8	1,300.3	0.0	48.3	2,157.9	0.0	1968.5	Contam.	1966.0	contaminated with CFC-12. Recharge date about
		130.4	1,398.4	ERR	50.0	2,320.7	ERR	1968.5	Contam.	ERR	25 years ago.
		127.4	1,340.5	0.0	48.9	2,224.7	0.0	1968.5	Contam.	1966.0	
WI Ce 204	11/9/93	65.5	107.5	11.1	25.1	178.5	10.3	1964.5	1972.5	1972.0	CFC-113 for samples number 2 and 4 are
		69.0	119.3	10.0	26.5	198.0	9.4	1965.0	1973.5	1971.0	estimates. Recharge date about 1970.
		102.9	103.5	10.2	39.5	171.9	9.6	1967.0	1972.5	1971.5	
WI Ce 210	11/24/93	137.5	127.7	9.8	52.7	211.9	9.1	1969.0	1974.5	1971.0	Recharge date early 1970's.
		136.8	137.7	11.2	52.5	228.6	10.4	1969.0	1975.0	1972.0	
		137.0	146.1	10.0	52.6	242.5	9.4	1969.0	1976.0	1971.0	
WI Ce 242	11/29/93	ERR	1,192.8	ERR	ERR	1,979.6	ERR	ERR	Contam.	ERR	Highly contaminated with CFC-12. The water
		ERR	1,258.0	ERR	ERR	2,087.7	ERR	ERR	Contam.	ERR	has very small amounts of CFC-11 and CFC-113.
		ERR	1,256.5	ERR	ERR	2,085.2	ERR	ERR	Contam.	ERR	May be more than 30 years old.
WI Ce 244	12/1/93	15,591.0	2,439.8	ERR	5,980.6	4,049.0	ERR	Contam.	Contam.	ERR	Highly contaminated with CFC-11 and CFC-12.
		15,177.1	2,331.5	ERR	5,821.5	3,869.2	ERR	Contam.	Contam.	ERR	Cannot determine CFC-113 because of
		15,589.5	2,414.9	ERR	5,979.7	4,007.7	ERR	Contam.	Contam.	ERR	interferences.



EXPLANATION

ESTIMATED TRITIUM CONCENTRATION IN PRECIPITATION AT WASHINGTON, D.C., DECAY CORRECTED TO NOVEMBER 1993. CURVE DERIVED FROM DATA CORRECTED AT OTTAWA, ONTARIO, CANADA.

TRITIUM CONCENTRATION IN PRECIPITATION AT WASHINGTON, D.C., DECAY CORRECTED TO NOVEMBER 1993

WI Ce 242 — TRITIUM CONCENTRATION IN GROUND-WATER SAMPLES FROM THE SALISBURY AQUIFER. NUMBER NEXT TO SYMBOL IS THE WELL NUMBER.

Figure 17. - Tritium concentrations in the Salisbury aquifer compared to decay-corrected tritium in precipitation at Washington, D.C.

Ontario (fig. 17). The pre-1959 tritium data in precipitation at the Washington, D.C. site was estimated using tritium concentrations collected at Ottawa, Ontario.

A summary of the estimated recharge dates using tritium and chlorofluorocarbons and the sampled well-screen lengths and positions are given in table 4. Two ranges of dates are possible for most samples: before the peak years of tritiated precipitation and after the peak years. The lowest concentration was sampled in observation well WI Ce 242, adjacent to production well 2 (WI Ce 241). This well is screened in the lower part of the paleochannel. The low tritium concentration may be a result of the well-screen depth and its proximity to the production well. The absence of a confining layer separating the Salisbury aquifer from the underlying Manokin aquifer, and the generally greater head in the Manokin aquifer relative to the Salisbury aquifer, may have resulted in mixing of water from the two aquifers. Pumping from the production well may increase the flow of

water upward from the Manokin aquifer, but most of the water flowing to the production well is from the Salisbury aquifer. The Manokin aquifer is confined throughout most of the study area, and the recharge area (subcrop area beneath the Pensauken Formation) is about 1.5 mi to the northeast; therefore, the Manokin aquifer water at this location is probably older than the Salisbury aquifer water. The estimated age for the sample from well WI Ce 242 is 1953-55, or a mixture of younger Salisbury aquifer water and pre-bomb Manokin aquifer water; however, considering that most of the water at this depth is from the Salisbury aquifer and that the CFC concentrations indicate a pre-1963 date, it is more likely that the 1953-55 estimate is correct.

The estimated recharge date for a water sample collected from well WI Ce 241 (Paleochannel production well 2) ranges between either 1954-61 or 1975-present. Because of the wide interval screened (105-195 ft below land surface), it is assumed that this sample is a mixture of Salisbury and Manokin aquifer water. Chlorofluorocarbons were not analyzed at this site.

The estimated recharge date of a water sample collected from WI Ce 200 (Paleochannel production well 1) ranges between either 1954-61 or 1975-present. Similar to production well 2, this well has a wide screen interval (80-160 ft below land surface) that allows mixing of water of different ages.

Water samples were collected from observation wells WI Ce 204 and WI Ce 210, at distances from production well 1 of approximately 45 and 600 ft, respectively. The estimated recharge dates range between either 1954-61 or 1972-present for well WI Ce 204 and between 1954-61 or 1968-92 for well WI Ce 210. These wells are screened at about the same depths below sea level--109 to 119 ft for well WI Ce 204 and 122 to 132 ft for well WI Ce 210. The latter estimated date ranges are more likely, because the CFC analyses resulted in estimated dates in the early 1970's for both wells.

The water sample collected from observation well WI Ce 196 yielded a possible recharge date ranging between 1954-61 or 1968-91. This well is located between Little Burnt Branch and well WI Ce 200 (production well 1) (fig. 2) and is screened from 6 to 7 ft below land surface. The sample was collected during a period of low precipitation and higher-than-average pumpage from production well 1. Baseflow and potentiomanometer measurements made 10 days after the tritium sample was collected indicate that the stream was losing water during this period, and the tritium date may represent infiltrating stream water mixing with native aquifer water. The relatively recent recharge date estimated by CFC-dating (early 1980's) also indicates mixing of water from the two sources.

Table 4. - Estimated recharge date of water from wells screened in the Salisbury aquifer using tritium and chlorofluorocarbons

[TU, tritium units; CFC, chlorofluorocarbon; --, not determined; NA, not analyzed]

Well number	Date	Tritium concentration (TU)	Possible tritium recharge date	CFC-model recharge date	Length of well screen (feet)	Well screen position (feet below land surface)
WI Ce 13	12/1/93	15.3 +/- 1.1	1954-61; 1968-91	1985	20	45 - 65 (?)
WI Ce 196	11/23/93	16.5 +/- 1.2	1954-61; 1968-91	1983	1	6 - 7
WI Ce 200	11/23/93	8.3 +/- 0.7	1954-61; 1975-present; see note A	1968	80	80 - 160
WI Ce 204	11/9/93	11.1 +/- 0.9	1954-61; 1972-present	1970	10	109 - 119
WI Ce 210	11/24/93	14.3 +/- 1.1	1954-61; 1968-92	early 1970's	10	122 - 132
WI Ce 241	11/29/93	8.4 +/- 0.7	1954-61; 1975-present; see note A	NA	90	105 - 195
WI Ce 242	11/29/93	2.2 +/- 0.4	1953-55; see note A	pre-1963	10	175 - 185
WI Ce 244	12/1/93	17.2 +/- 1.3	1954-61; 1968-91	--	16	38 - 54

A -- Mixture of older, pre-bomb Manokin aquifer water and younger Salisbury aquifer water.

The wells sampled for tritium in the Park well field include WI Ce 13 and WI Ce 244 (production well 2a). The possible recharge-date ranges for both wells is estimated to be either 1954-61 or 1968-91. Given the shallow screen depth of production well 2a (38-54 ft below land surface), it is surprising that the recharge date is not closer to 1993. The older recharge dates may suggest more horizontal flow from distant recharge areas as opposed to vertical flow from areas adjacent to the well.

SIMULATION OF GROUND-WATER FLOW

MODEL DESCRIPTION

Ground-water flow in the Salisbury and Manokin aquifers within the Upper Wicomico River Basin was simulated by use of the U.S. Geological Survey's digital, three-dimensional, finite-difference model, MODFLOW (McDonald and Harbaugh, 1988). The model simulated steady-state conditions using an estimated 1993 pumpage distribution. Model geometry was developed based on the hydrogeologic framework discussed earlier in the report. Starting heads used in the model approximated the median water-table surfaces shown in figures 9 and 12. The input parameters used in the model included recharge, horizontal hydraulic conductivity, transmissivity, vertical leakance, river stage, river-bottom elevation, and river-bottom hydraulic conductance, and pumpage. The model was calibrated to median water levels measured in the Salisbury aquifer by adjusting recharge, horizontal hydraulic conductivity in the Salisbury aquifer, and river-bottom hydraulic conductivity. Simulated baseflow in the model was checked against the median baseflow in Beaverdam Creek, calculated from the period of record 1931-76, and the estimated baseflow in eight other streams within the study area.

CONCEPTUAL MODEL

A conceptual model of the ground-water-flow system (fig. 18) was developed on the basis of the hydrogeologic framework and aquifer properties described in this report. Important aspects of the natural flow system incorporated into the conceptual model include the direction of ground-water flow (both laterally and vertically), physical boundaries (rivers and impermeable boundaries), and flow boundaries (ground-water-flow divides).

The ground-water basin of the Salisbury aquifer in the Salisbury area was assumed to correspond to the surface-water basin of the Upper Wicomico River. In general, water enters the Salisbury aquifer in the Upper Wicomico River Basin from direct precipitation. The ground water discharges to small streams, ponds, the Wicomico River, and wells. In the vicinity of the paleochannel, deeper ground-water flow reaches the city's pumped wells. Water also flows into the paleochannel from the deeper Manokin aquifer. Away from the paleochannel, ground-water flow in the Manokin aquifer is generally to the south.

Some water likely moves vertically upwards from the Manokin aquifer to the Salisbury aquifer through the confining bed. The exchange of water between the Manokin and the Salisbury aquifers, however, is probably small considering the low relative head gradients and low permeability of the intervening confining bed (clay). It is assumed that flow does not occur across the St. Marys Formation (confining bed), because of its thickness (from 77 to 93 ft) and low permeability.

In most of the topographically high areas within the basin, the unsaturated zone of the Salisbury aquifer is overlain by the lower permeability Walston Silt. This unit was not included in the model, because it generally lies above the water table and is not areally extensive.

The ground-water system is considered to be in a state of dynamic equilibrium. Hydraulic heads in the Salisbury aquifer fluctuate with the seasons on the order of 1 to 6 ft, but, when viewed over a period of several years, the water-level change is negligible (fig. 10). Also, water levels within the Salisbury aquifer are not greatly affected by pumpage (figs. 10 and 11). Few data are available on the head change in the Manokin aquifer, because of the absence of observation wells in that aquifer. Because there are no significant withdrawals from the Manokin aquifer within the study area, it is assumed that the altitude of the potentiometric surface is roughly the same as was mapped by Boggess and Heidel (1968).

GRID DESIGN AND MODEL LAYERS

The model grid consists of five layers, 80 columns, and 84 rows and covers the entire Upper Wicomico River Basin (fig. 19). The grid is oriented along latitudinal lines and measures 10.6 mi in the north-south direction and 9.5 mi in the east-west direction. The total area covered by the grid is approximately 100 mi^2. The grid spacing is finest in the areas of the Paleochannel well field and the Park well field. Cell dimensions at those sites are 100 x 100 ft and 150 x 150 ft, respectively. The number of cells in each model layer is 6,720 and the total number of cells in the model is 33,600.

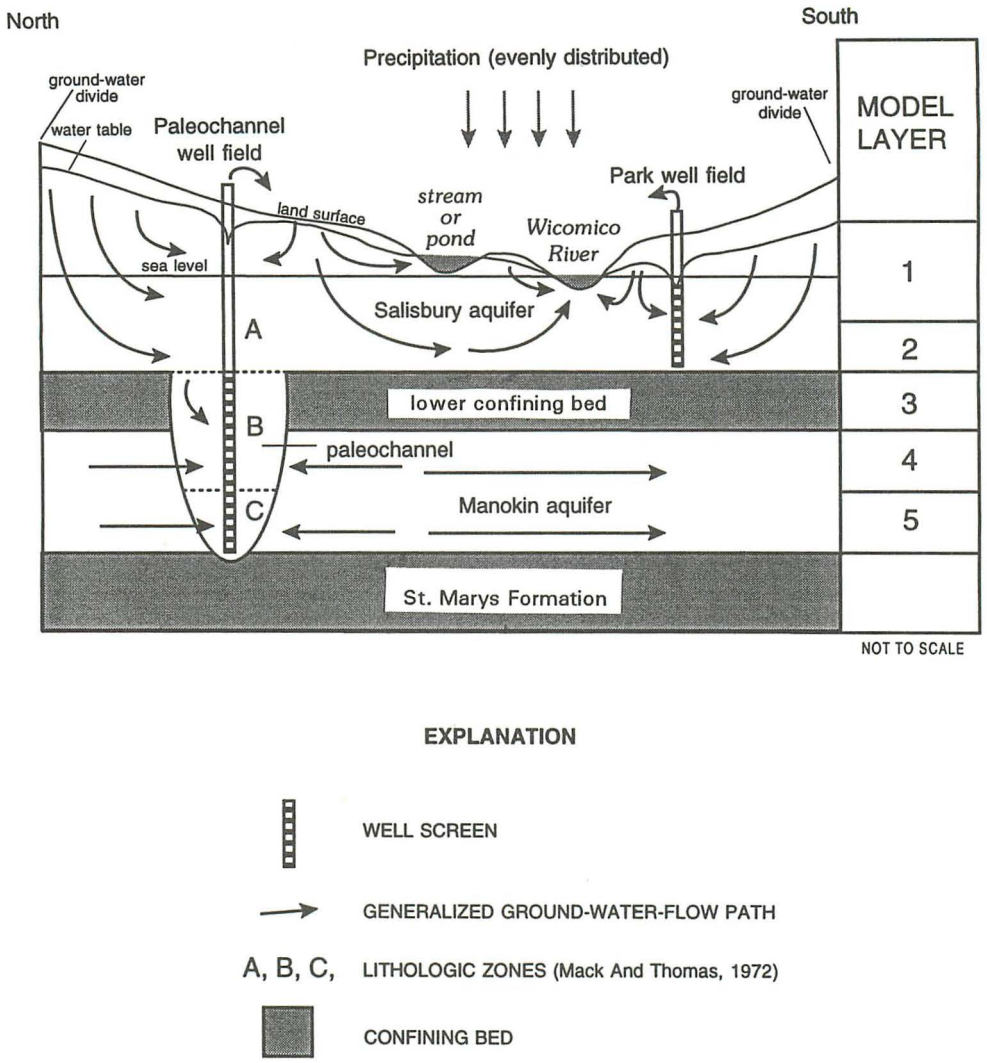


Figure 18. - Conceptual model of the ground-water-flow system within the boundaries of the Upper Wicomico River Basin.

The layers of the model generally correspond to the natural hydrogeologic units within the study area (tab. 1 and fig. 18). The top layer represents the aquifer characteristics of the Beaverdam Sand extending from the water-table surface to about 10 to 15 ft below sea level. The Parsonsburg Sand and the Walston Silt are not included in this layer, because they are relatively thin within the model area or are generally above the saturated zone.

Layer 2 represents the upper part of the Pensuaken Formation (or red gravelly facies). Pumpage from the Park well field, along with most of the ground-water use over 10,000 gal/d, occurs within this layer. The part of the Pensauken above the paleochannel and the Beaverdam Sand were described as one separate hydrogeologic unit (Zone A) by Mack and Thomas (1972) in the area of the paleochannel at Naylor Mill Road. The lithology of Zone A is a fairly uniform, coarse brown sand interbedded with thin layers of clay. The unit acts as one hydrologic unit throughout the Salisbury area outside of the paleochannel and was divided in this study based on the differences in permeability between the upper and lower sections.

The third layer in the model represents the upper part of Zone B of Mack and Thomas (1972) in the area of the paleochannel, and the "lower confining bed" in areas outside the paleochannel. This layer includes the hydraulic properties of the confining bed separating the water-table aquifer from the Manokin aquifer and also, in places, the coarse gravelly sands found in the paleochannel.

The fourth layer represents, (1) within the paleochannel, the part of Zone B that lies at an altitude below the "lower

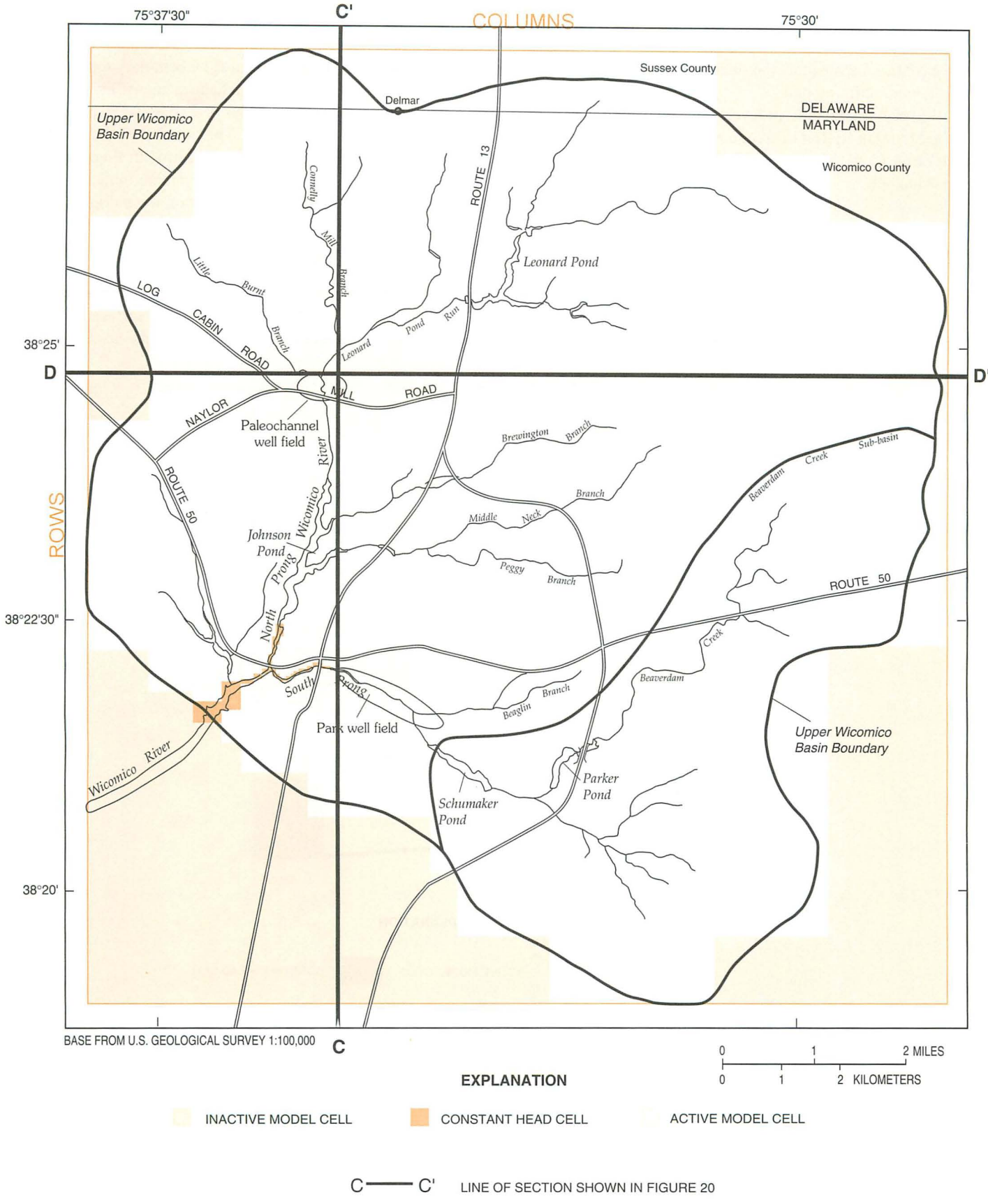


Figure 19. - Finite-difference, ground-water-flow model grid (model layers 1 and 2).

confining bed", and (2) the upper part of the Manokin aquifer outside the paleochannel. The thickness of this layer was set at 55 ft, which equals the thickness of Zone B beneath the "lower confining bed."

Layer five, the bottom layer, represents the lower part of the Manokin aquifer in areas outside of the paleochannel, and Zone C of Mack and Thomas (1972) within the paleochannel.

BOUNDARY CONDITIONS

The lateral boundaries used in the flow model correspond to both natural flow boundaries and physical boundaries. Cross sections along model row 25 and column 35 illustrate the relative position and types of boundaries used in the model (fig. 20). The perimeter of the Upper Wicomico River Basin was chosen as a no-flow boundary for layers 1 and 2 (Salisbury aquifer). In the shallow, unconfined ground-water system, the surface-water drainage divide coincides with the ground-water divide. The position of the ground-water divide is assumed to be constant with time and not affected by pumping wells. Any movement in the ground-water divide would have little effect on the model results, because it is so far from the largest stresses in the system (streams and large well fields). The choice of boundary was tested by simulating a constant-head boundary in place of the no-flow boundary, and specifying heads in the boundary cells as the altitude of the water-table surface. Results of the simulation were similar to those of the no-flow con-

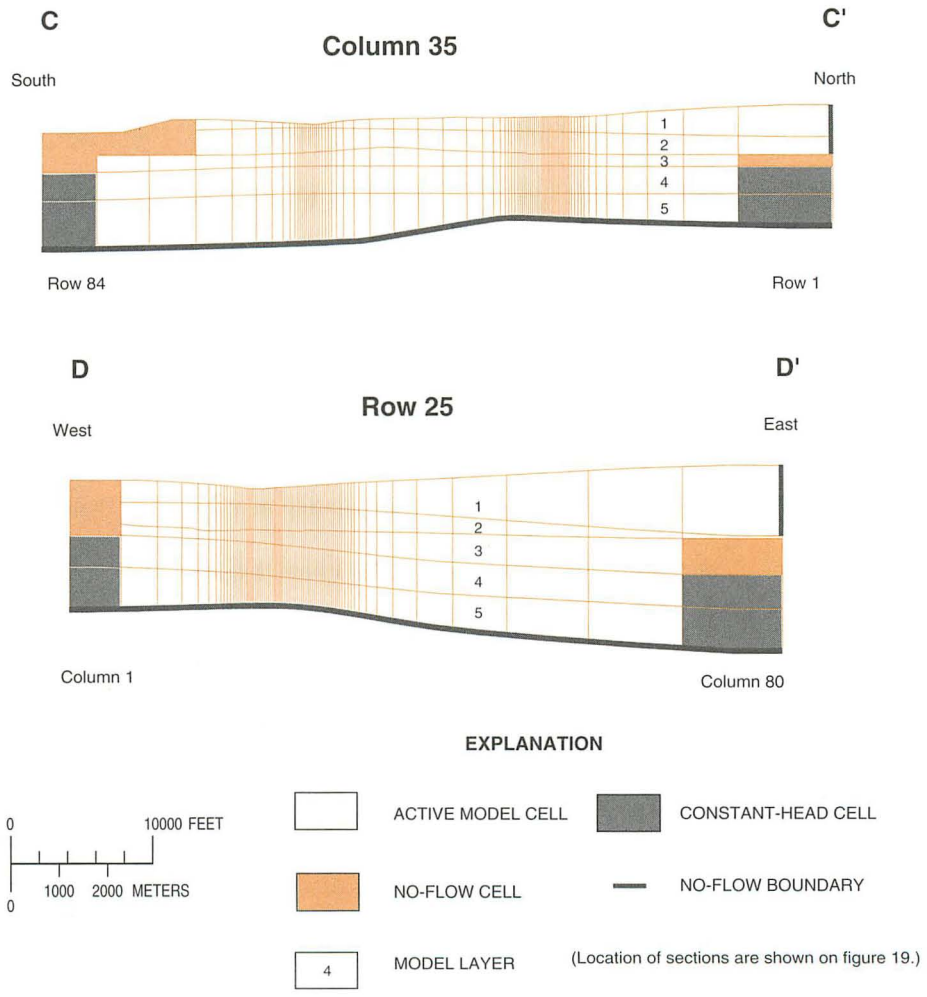


Figure 20. - Cross sections along model row 25 and column 35 showing model grid, layers, and boundaries.

dition, suggesting that the choice of boundary type would not significantly change the results of the simulations described in this report. A constant head (sea level) boundary exists along the tidal portion of the Wicomico River and was represented in both layers 1 and 2 (fig. 19).

The lateral boundary of layer 3 was set as no-flow along the perimeter of the model grid. Layer 3 represents a low-permeability confining bed in all areas outside the paleochannel. Horizontal flow in the paleochannel beneath the basin divide is negligible.

Because the Manokin aquifer (model layers 4 and 5) extends beyond the Upper Wicomico River Basin, it has no natural physical or flow boundaries within the study area. A constant-head boundary was used at the model edges to approximate the estimated heads in those areas (fig. 21). The head distribution along the west and northwest perimeter of the model was estimated from ranges of water-table altitude in the recharge area of the Manokin aquifer (Boggess and Heidel, 1968, fig. 7, p. 13). Heads along the boundary range from 20 to 45 ft above sea level. The constant-head boundary was tested by replacing it with a no-flow boundary, and no appreciable differences in results were evident.

The base of the Manokin aquifer was represented by a no-flow boundary, because the underlying St. Marys Formation is a relatively thick (80-90 ft), low-permeability clay. Leakage through this unit is insignificant for the objectives of this study.

HYDRAULIC PARAMETERS

Horizontal Hydraulic Conductivity and Transmissivity

Hydraulic conductivity arrays are required by MODFLOW to simulate unconfined aquifers. Simulation of confined aquifers requires transmissivity arrays. In this model, layer 1 was designated as unconfined and layers 2, 3, 4, and 5 designated as confined. Transmissivity arrays were calculated by multiplying horizontal hydraulic conductivity by the model-layer thickness in each model cell. Horizontal hydraulic conductivity of the hydrogeologic units is discussed in the geohydrology section of the report.

Horizontal hydraulic conductivity of layer 1 was initially specified as 200 ft/d throughout the model area. Preliminary model simulations resulted in heads too low in the west-central part of the basin. To correct this problem, the horizontal hydraulic conductivity was lowered to 100 ft/d in that area. Slug tests performed in wells screened in the upper part of the Salisbury aquifer at the Newland Park landfill showed an average horizontal hydraulic conductivity of 76 ft/d (GeoSyntec Consultants, 1992).

The transmissivity of model layer 2 was initially calculated using a horizontal hydraulic conductivity value of 300 ft/d. This value was lowered to 100 ft/d in the west-central part of the basin to correct for low simulated heads. The variation in transmissivity across the study area is shown in map view in figure 22 and in sectional view along column 35 in figure 23. The transmissivity of layer 2 does not exceed 10,000 ft^2/d in the western part of the model, because of the lower horizontal hydraulic conductivity in that area and on the east side of the model area where eastward thinning of the Pensauken Formation occurs.

Transmissivity in the paleochannel area of model layer 3 was calculated using a horizontal hydraulic conductivity of 670 ft/d. The area of high transmissivity (5,000 to 15,000 ft^2/d) generally corresponds to the location of the paleochannel as mapped by Weigle (1972) (fig. 22). Outside the paleochannel ("lower confining bed") in layer 3, the transmissivity is less than 1 ft^2/d. Transmissivity of the confining bed was calculated using a horizontal hydraulic conductivity value of 1.0×10^{-2} ft/d. This value falls between those typical of silt (0.26 ft/d) and clay (7.0×10^{-4}) (Todd, 1980). It was arrived at using the vertical hydraulic conductivity of the confining bed and a vertical to horizontal hydraulic conductivity ratio of 1:100. To simulate the transition from confining layer to aquifer in layer 3, the hydraulic conductivity was increased gradually over several model nodes across the paleochannel boundary.

Transmissivity of layer 4 ranges from 2,000 to 6,000 ft^2/d outside the paleochannel area and 20,000 to 31,000 ft^2/d in the paleochannel. Transmissivity values were calculated based on horizontal hydraulic conductivity values of 54 ft/d outside the paleochannel and 670 ft/d in the paleochannel.

Transmissivity of layer 5 ranges from 1,500 to 3,500 ft^2/d in the north-central part of the model area to 3,500 to 5,500 ft^2/d in other locations (fig. 22). The variation in transmissivity is caused by changes in the estimated thickness of the aquifer. A horizontal hydraulic conductivity of 54 ft/d was used to calculate transmissivity.

No aquifer test data are available for Zone C (layer 5 within the paleochannel). However, Mack and Thomas (1972) indicate that the lithology consists of fine to coarse sand capable of yielding moderate amounts of water. A hydraulic conductivity equal to that of the Manokin aquifer was assigned initially to this unit.

Vertical Leakance

Vertical leakance is a mathematical expression used to calculate the volume of flow of water between model layers. Vertical flow (L) is defined as the vertical hydraulic conductivity (K_v) of a layer divided by the thickness of the layer (b):

$$L = \frac{K_v}{b}$$

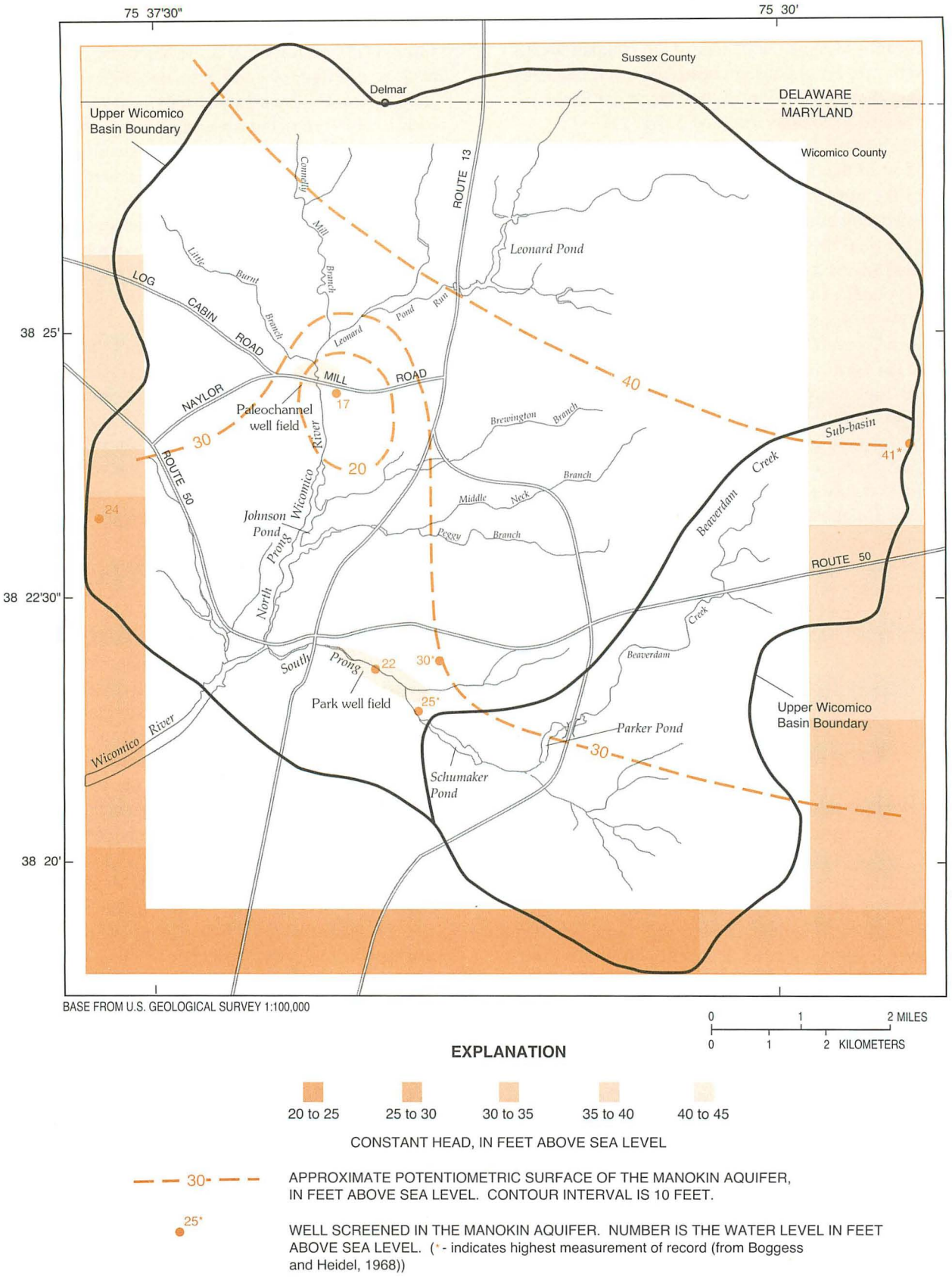
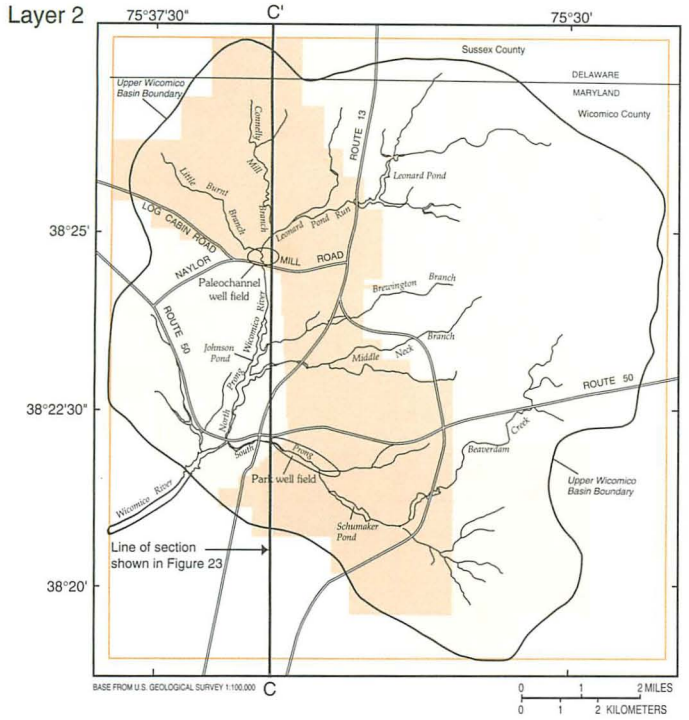


Figure 21. - Approximate potentiometric surface of the Manokin aquifer and values assigned to constant - head model cells along boundary of layers 4 and 5.

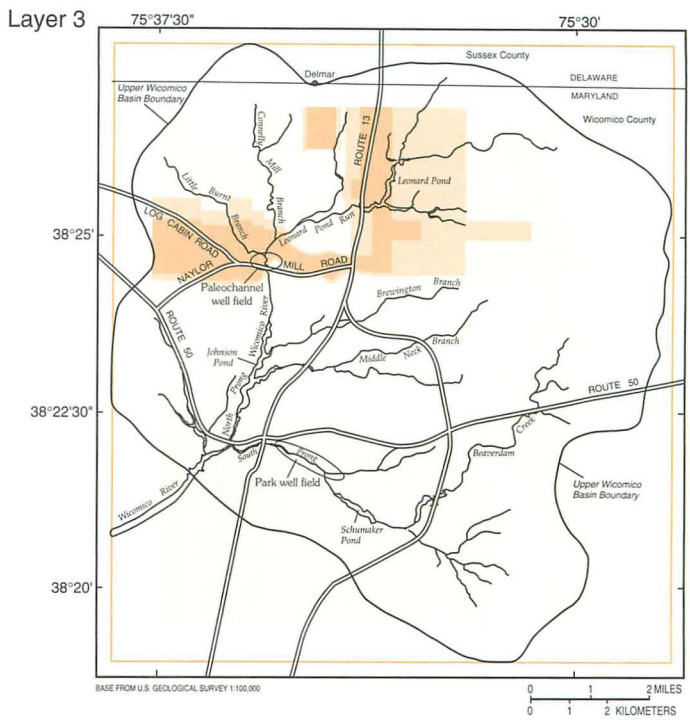


TRANSMISSIVITY

UP TO 10,000 FT²/D

10,000 TO 20,000 FT²/D

AREA OF NO-FLOW



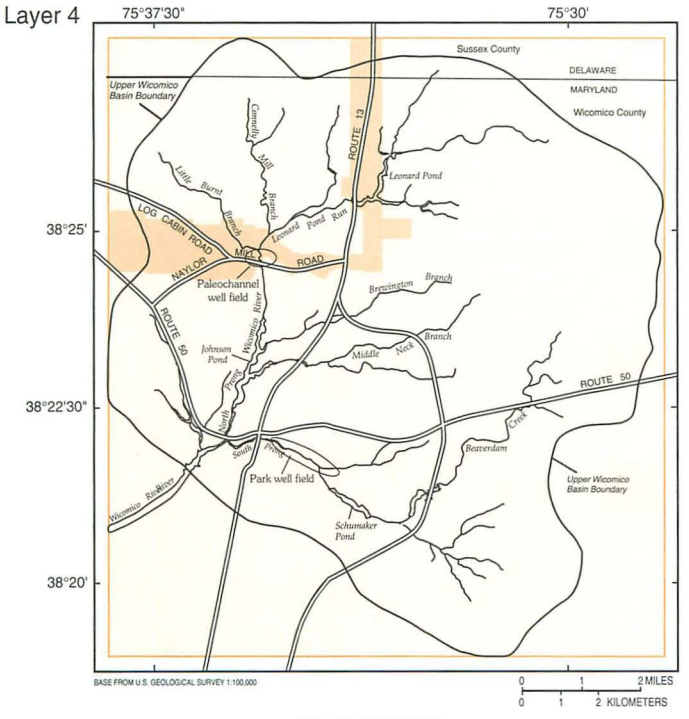
TRANSMISSIVITY

UP TO 1 FT²/D

1 TO 5,000 FT²/D

5,000 TO 15,000 FT²/D

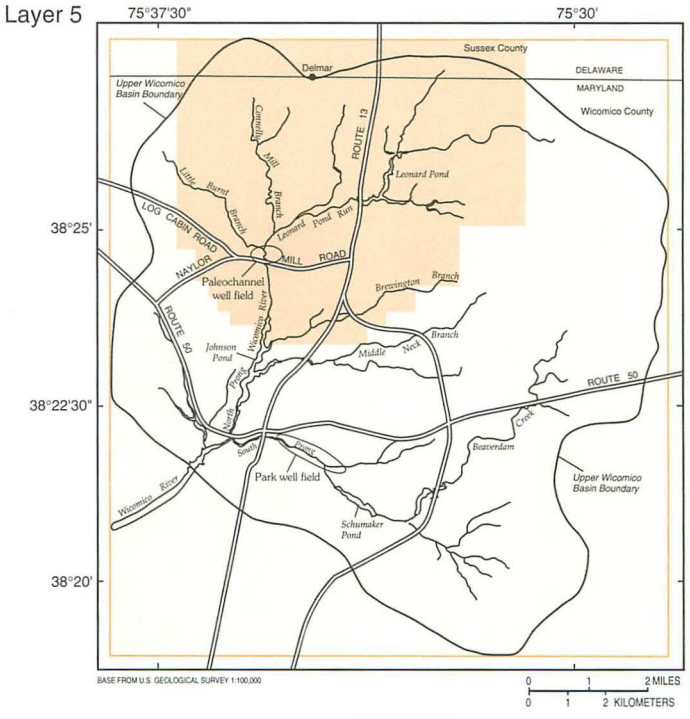
AREA OF NO-FLOW



TRANSMISSIVITY

2,000 TO 6,000 FT²/D

20,000 TO 31,000 FT²/D

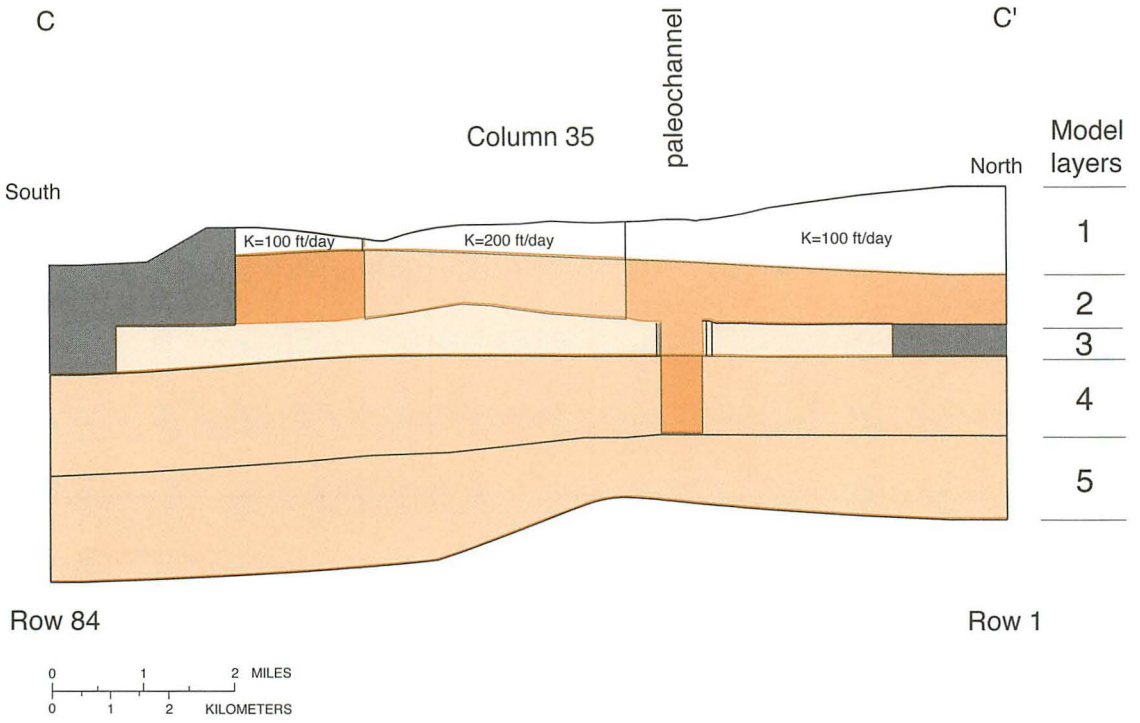


TRANSMISSIVITY

1,500 TO 3,500 FT²/D

3,500 TO 5,500 FT²/D

Figure 22. - Layers showing transmissivity zones used in calibrated ground-water flow model.



EXPLANATION

TRANSMISSIVITY, ft^2/d

AREA OF NO-FLOW

K — VARIABLE TRANSMISSIVITY. HORIZONTAL HYDRAULIC CONDUCTIVITY = 100 AND 200 FT/DAY

0 TO 1

1,500 TO 5,500

6,000 TO 15,000

15,000 TO 31,000

(Location of section is shown in figure 22)

Figure 23.- Cross section showing ranges of transmissivity along model column 35.

Vertical leakance terms used in the model represent two different cases—sand-on-sand contacts and sand-on-clay contacts. In the case of the sand-on-sand contacts, the vertical hydraulic conductivity was estimated at 1/10th the horizontal hydraulic conductivity. Vertical leakance between model layers were calculated for each model cell using the following equation (Chiang and Kinzelbach, 1993):

$$\frac{1}{L} = \frac{(b_1)/2}{K_1} + \frac{(b_2)/2}{K_2},$$

where

L = vertical leakance, d^{-1};
K_1 = vertical hydraulic conductivity of the layer above, ft/d;
K_2 = vertical hydraulic conductivity of the layer below, ft/d;
b_1 = thickness of the upper layer, ft; and
b_2 = thickness of the lower layer, ft.

Leakance between model layers 1 and 2 represents a sand-on-sand contact. Vertical hydraulic conductivity used to calculate leakance was 20 ft/d. Leakance between layers in the paleochannel represents a sand-on-sand contact. It was calculated in each case using the estimated vertical hydraulic conductivity of the overlying layer. Leakance between layers 2 and 3 outside the paleochannel represents a sand-on-clay contact. A vertical hydraulic conductivity of 0.001 ft/d was used to calculate the leakance. This value represents the vertical hydraulic conductivity of the "lower" confining bed as determined by field and laboratory tests performed near the Park well field (Wolff, 1970). Leakance between layers 3 and 4--also representing a sand-on-clay contact--was calculated using the same vertical hydraulic conductivity value of 0.001 ft/d. Leakance was increased gradually in model cells at the edges of the paleochannel to simulate the transition between sand-on-clay and sand-on-sand (paleochannel area) conditions. Leakance between layers 4 and 5 outside the paleochannel represents a sand-on-sand contact and was calculated using a vertical hydraulic conductivity of 5.4 ft/d.

RECHARGE

An initial ground-water recharge rate of 13.4 in/yr (0.0031 ft/d) was applied evenly to model layer 1 (Salisbury aquifer). The total active surface area of model layer 1 is approximately 2.1 x 10^9 ft^2; therefore, the total volume of water entered as recharge into the model was 6.5 x 10^6 ft^3/d. During model calibration, this value was adjusted upward to 14.0 in/yr (0.0032 ft/d) or 6.7 x 10^6 ft^3/d over model layer 1.

PUMPAGE

Ground-water pumpage in the model was simulated as constant flux boundaries in cells that contain one or more pumping wells. All major (over 10,000 gal/d) well appropriations in 1993, including public and domestic supply, industry, and irrigation, were simulated in the model. Pumpage was assigned to model cells based on well location and well-screen depth. Total pumpage from the Salisbury aquifer was 8.7 Mgal/d in 1993. Domestic pumpage was placed in model layer 2 based on the assumption that most of the domestic pumping was from the lower part of the Salisbury aquifer.

The distribution of all major pumpage in the Salisbury aquifer in model layer 2 is shown in figure 24. The northeast and southeast parts of the model area, and areas between cells showing pumpage, contain a few wells supplying homes and farms. Because the amount of water withdrawn in these areas is relatively small, it was not considered necessary to include these wells in the model simulations. Pumpage within the model area is divided into three groups, (1) less than 10,000 gal/d, (2) 10,000 to 100,000 gal/d, and (3) 100,000 to 1,000,000 gal/d.

The only pumpage from model layer 3 was the upper part of the well screen in Paleochannel well 1 (0.97 Mgal/d). Pumpage from model layer 4 consisted of the lower part of the well screen in Paleochannel well 1 (2.5 Mgal/d) and the Town of Delmar supply (0.97 Mgal/d).

In the calibrated model, pumpage in the Paleochannel well field consisted of 3.5 Mgal/d from WI Ce 200 (well 1). This well was pumped because it is used more often than Paleochannel well 2 (WI Ce 241). In particle-tracking simulations, Paleochannel wells 1 and 2 were pumped separately at the rate of 3.5 Mgal/d (1993 pumpage) to determine the contributing areas for those wells. Pumpage from the Park well field totalled 1.7 Mgal/d and was divided evenly over wells WI Ce 244 (2a), Ce 246 (6a), Ce 247 (7b), Ce 248 (8a), Cf 194 (10a), Cf 183 (14a), and Cf 195 (16a) at a rate of 0.24 Mgal/d.

Pumpage from the Manokin aquifer was insignificant in relation to pumpage in other layers, and was not simulated.

STREAM SIMULATION

Surface-water bodies (streams, non-tidal rivers, and ponds) were simulated in the model as non-penetrating streams using the river package of MODFLOW. The length, width, and bottom thickness of each river reach contained in a model cell was input into the river package. During model simulations, flux between river cells and model layer 1 is controlled by the head gradient established between the river cells and model layer 1 and the hydraulic conductivity of the river-bottom material. An estimate of the hydraulic conductivity of the river-bottom material is multiplied by the area of the river bottom, divided by the streambed thick-

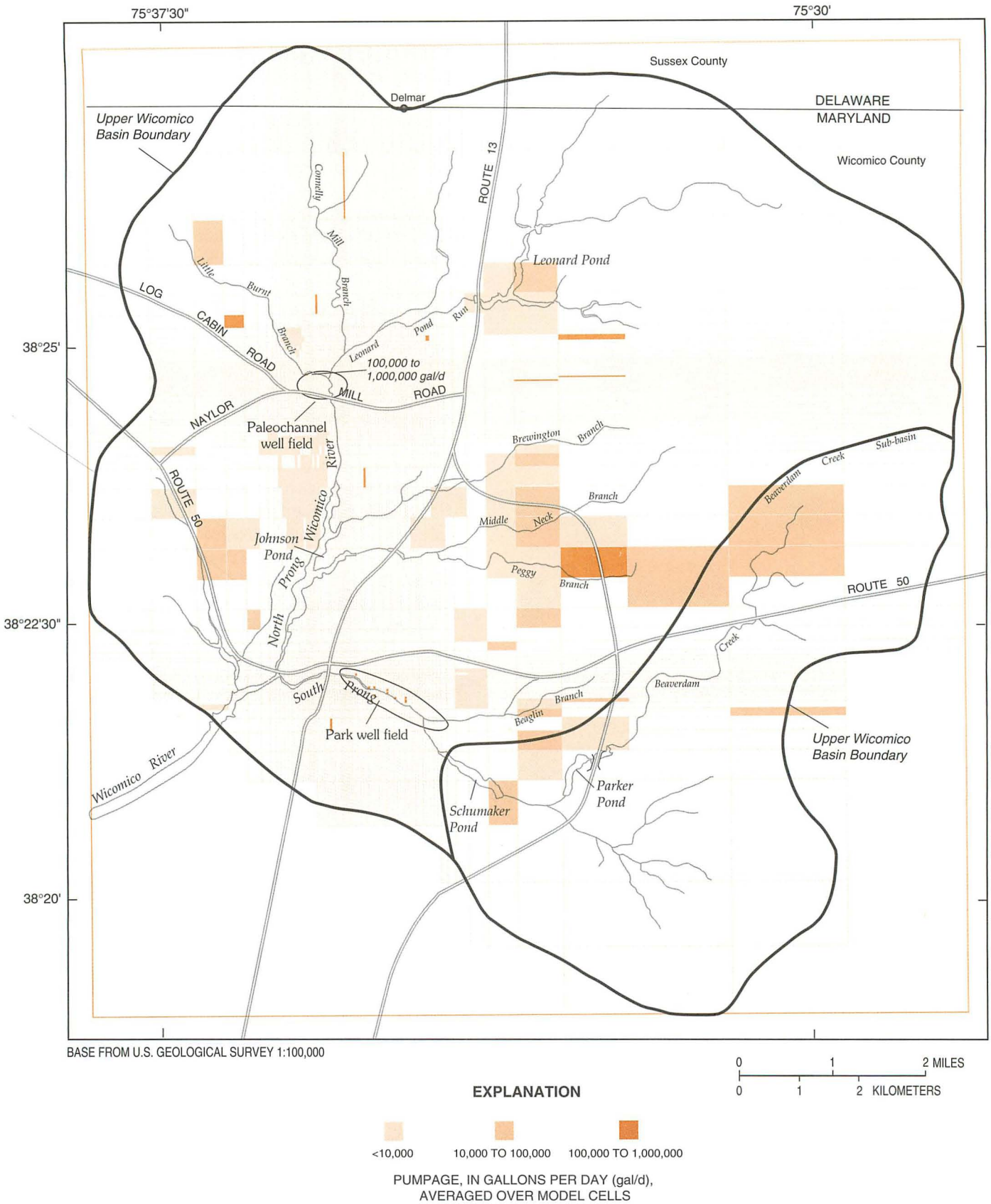


Figure 24. - Pumpage used in the calibrated steady-state model in the Salisbury aquifer (model layer 2).

ness, and the resulting hydraulic conductance term is assigned to each river cell. For this model, the hydraulic conductivity of the river-bottom material was set at 3.75 ft/d, which is within the range of values for a silty to clean sand (Freeze and Cherry, 1979). Average river stage and bottom altitudes were also assigned for each river cell. The river stage was estimated at 1 ft above the stream bottom and remained constant throughout the simulation. River-bottom altitudes and lengths of river reaches contained in the corresponding model cells were assigned from 7.5-minute topographic quadrangles for the perennial streams within the basin. Near the two Salisbury well fields, river-bottom altitudes were estimated using 1993 orthophotographic maps at a scale of 1:1,200 (1 in. = 100 ft) supplied by the Department of Public Works, City of Salisbury.

The tidal portion of the Wicomico River was simulated as a constant-head boundary with the head set to mean sea level. Cells in both layers 1 and 2 under the river were set to constant head, because there is probably a good hydraulic connection between those layers.

STEADY-STATE CALIBRATION

Early model simulations indicated that the flow model was most sensitive to recharge, horizontal hydraulic conductivity of layer 1, transmissivity of layer 2, and river-bottom hydraulic conductance. Model calibration was achieved by adjusting these parameters until simulated heads and budget terms approximated medians of observed heads and baseflow. Recharge, horizontal hydraulic conductivity, and riverbed conductance were systematically adjusted over realistic ranges, and the sum of the absolute differences between measured and simulated heads in the Salisbury aquifer and the hydrologic budgets were compared. The changes made in the input parameters were within ranges considered reasonable based on available field data.

It is assumed that the ground-water system is in a state of dynamic equilibrium. Therefore, simulated heads were compared to the median heads obtained from four wells with long-term water-level records and from 61 wells with either a few measured levels or with less than 2 years of continuous water-level record (appendix C). The median water levels are assumed to represent approximate steady-state hydrologic conditions within the basin. Because of non-systematic collection of these water levels, however, they may not represent true steady-state conditions.

Graphs of the sums of absolute differences between simulated and measured heads in the Salisbury aquifer were used to find the calibration point of the model (fig. 25). Water-level data from 66 wells were used in the calibration. River-bottom hydraulic conductivity was varied from 1.0 to 6.6 ft/d while horizontal hydraulic conductivity in layers 1 and 2 were varied from 102 to 306 ft/d and 157 to 470 ft/d, respectively. Recharge was varied from 8.75 in/yr to 17.5 in/yr, while horizontal hydraulic conductivity in layers 1 and 2 was varied by the same amounts as above. Closed contours of the absolute differences between simulated and measured heads are positioned roughly in the center of both plots--river-bottom hydraulic conductivity of 3.75 ft/d, recharge of 14 in/yr, and horizontal hydraulic conductivity for layers 1 and 2 of 204 and 313 ft/d, respectively. The least amount of absolute difference between measured and simulated heads was 181 ft.

A map of the simulated water-table altitude and water-table measurements is shown in figure 26. The greatest difference between measured and simulated heads generally occurs with the higher water levels. This is a result of averaging of head in the larger model cells, which occurs in the upgradient areas of the model. Table 5 lists the well number, median water level measured in the well, and the simulated head obtained in the cell representing the well. Hydrologic parameters used in the calibrated steady-state flow model are listed in table 6.

Within the study area there is very little head data for the Manokin aquifer to use for model calibration. Simulated heads were compared to two water levels measured in wells near the Paleochannel and Park well fields (tab. 7). A third water level (24 ft above sea level), measured in well WI Ce 148, was located within the area of the constant head boundary (fig. 27). There is good agreement between the measured and simulated heads in both wells. In general, the simulated head distribution compares favorably with the estimated potentiometric surface (fig. 12). Because of the low permeability clay layer overlying the Manokin aquifer throughout most of the study area, an imperfect head match should have little effect on conditions in the overlying Salisbury aquifer and on the overall results of this study. A cone of depression centered around the Paleochannel well field developed in the simulated potentiometric surface of the Manokin aquifer. This was a result of leakage of water upwards from the Manokin to the Salisbury aquifer. Approximately 6 Mgal/d of water moved from the Manokin aquifer to the Salisbury aquifer in the model simulation.

Outflow to river cells in the model approximated median baseflow in Beaverdam Creek, and estimated annual average baseflow calculated from the sub-basin area in eight additional streams within the Upper Wicomico River Basin (U.S. Geological Survey, 1964). Baseflow was separated from long-term stream-flow records (1931-76) for Beaverdam Creek as discussed previously. The calculated value was about 19.3 ft^3/s or 1.7×10^6 ft^3/d for Beaverdam Creek, and the simulated value is 15.7 ft^3/s or 1.4×10^6 ft^3/d. Baseflow was estimated in the eight additional streams by multiplying the drainage area by an estimated average annual recharge rate of 14 in/yr. Table 8 shows the estimated average annual baseflow and simulated base flow in the eight streams. The location of the streams are shown on figure 1.

(Text continued on p. 49.)

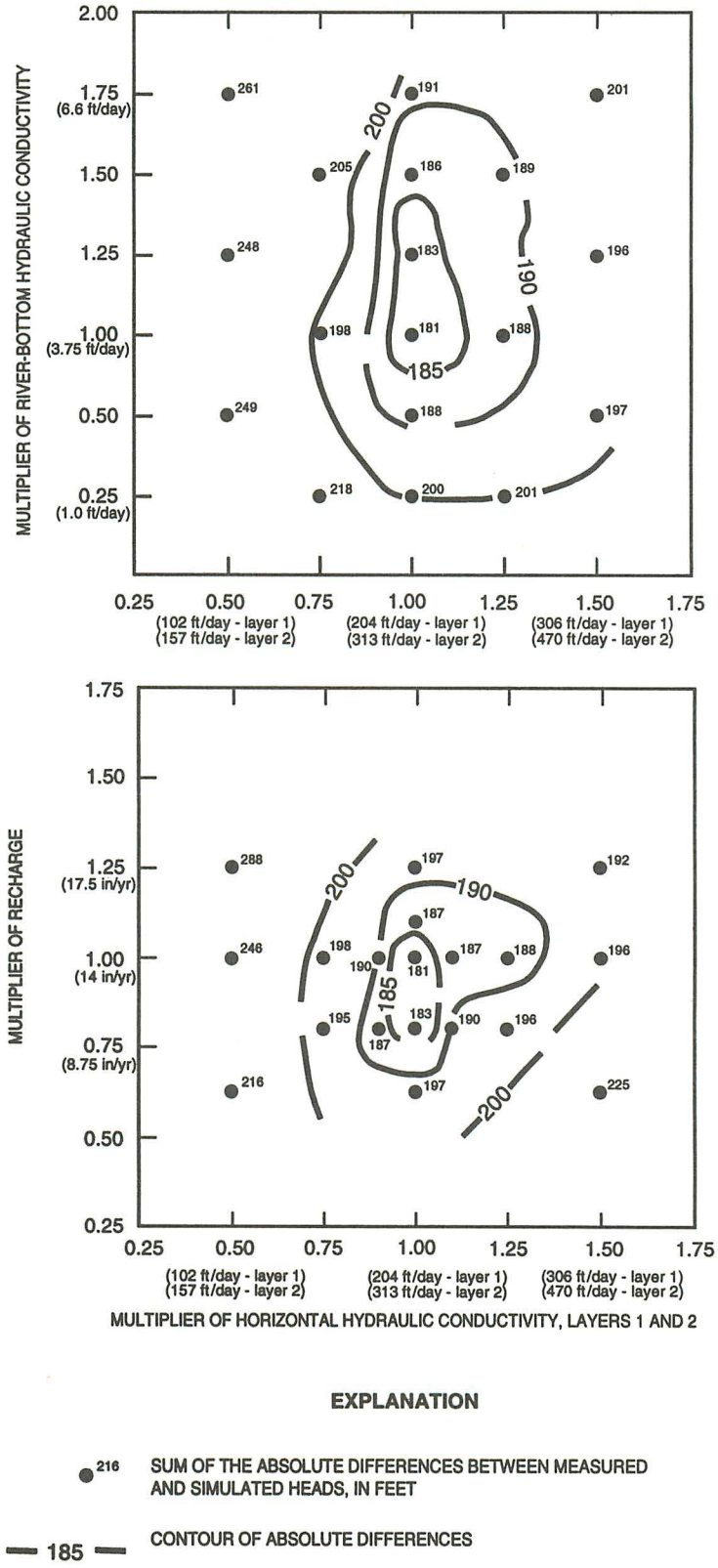
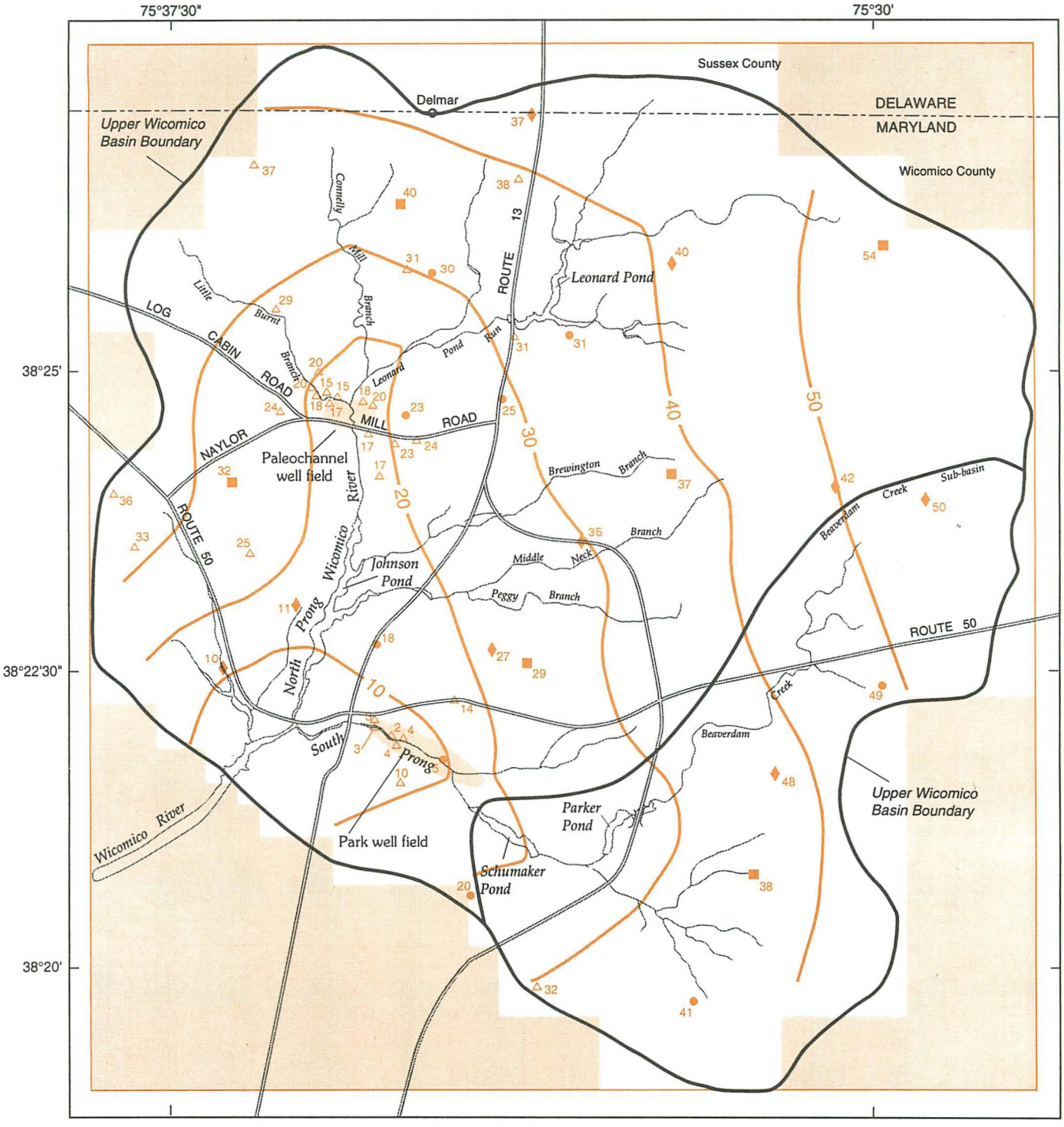


Figure 25. - Steady-state ground-water-flow model calibration graphs.



Base from U.S. Geological Survey, 1:100,000

0 1 2 MILES
0 1 2 KILOMETERS

EXPLANATION

10 — SIMULATED HEAD, IN FEET ABOVE SEA LEVEL. Contour interval is ten feet.

25 — WELL CONTROL POINT. Number next to symbol is the median water level for period of record (see appendix C.), in feet above sea level. Symbol type indicates the difference between simulated and median head, in feet: △ <2 ● 2-3 ■ 3-4 ♦ 4-16

Figure 26. - Simulated water table, measured water table, and calibration data points in the Salisbury aquifer.

Table 5. - Comparison of measured and simulated heads in the Salisbury aquifer (model layer 2) used in model calibration

[ft, feet; asl, above sea level]

Well number	Median measured head (ft asl)	Simulated head in cell representing well (ft asl)	Simulated head minus median head (ft)	Well number	Median measured head (ft asl)	Simulated head in cell representing well (ft asl)	Simulated head minus median head (ft)
WI Be 22	39.8	36.3	-3.5	WI Ce 243	20.2	18.6	-1.6
WI Be 52	36.7	35.3	-1.4	WI Ce 246	1.8	1.8	0
WI Be 56	29.3	30.8	+1.5	WI Ce 248	4.1	2.2	-1.9
WI Bf 6	40.2	45.4	+5.2	WI Ce 249	2.5	2.2	-0.3
WI Bf 20	30.1	32.7	+2.6	WI Ce 250	16.6	18.7	+2.1
WI Bf 21	30.6	31.7	+1.1	WI Ce 251	16.9	17.7	+.8
WI Bf 44	31.1	34.0	+2.9	WI Ce 252	23.5	21.1	-2.4
WI Bf 64	37.7	44.0	+6.3	WI Ce 253	23.9	23.2	-.7
WI Bf 66	38.3	39.0	+0.7	WI Ce 254	24.5	23.2	-1.3
WI Bf 69	31.0	31.3	+ .3	WI Ce 255	36.1	33.6	-2.5
WI Bg 3	53.8	57.9	+4.1	WI Ce 256	20.2	20.1	+.1
WI Ce 13	3.0	3.0	0	WI Cf 3	38.0	34.2	-3.8
WI Ce 16	4.5	4.0	-.5	WI Cf 17	37.4	41.1	+3.7
WI Ce 87	32.0	28.3	-3.7	WI Cf 52	41.6	50.6	+9.0
WI Ce 118	7.5	15.2	+7.7	WI Cf 81	24.9	28.4	+3.5
WI Ce 129	18.0	14.6	-3.4	WI Cf 95	27.2	21.9	-5.3
WI Ce 130	10.6	16.3	+5.7	WI Cf 102	29.0	25.5	-3.5
WI Ce 166	25.3	25.0	-.3	WI Cf 104	40.5	38.8	-1.7
WI Ce 174	10.5	9.7	- .8	WI Cf 105	36.0	32.0	-4.0
WI Ce 195	15.8	16.2	+ .4	WI Cf 107	28.1	24.8	-3.3
WI Ce 196	15.7	16.4	+ .7	WI Cf 112	48.5	32.6	-15.9
WI Ce 197	16.0	16.6	+ .6	WI Cf 117	14.1	16.5	+2.4
WI Ce 204	15.7	14.6	-1.1	WI Cf 147	24.1	22.6	-1.5
WI Ce 205	18.0	18.1	+.1	WI Cf 187	35.0	27.8	-7.2
WI Ce 206	15.2	16.4	+1.2	WI Cf 188	34.7	38.8	+4.1
WI Ce 207	15.1	16.4	+1.3	WI Cf 189	20.0	24.1	+4.1
WI Ce 208	20.3	19.6	-.7	WI Cf 195	5.1	8.0	+2.9
WI Ce 210	15.3	16.3	+1.0	WI Cf 196	23.2	22.1	-1.1
WI Ce 211	17.6	18.1	+.5	WI Cg 22	50.2	56.5	+6.3
WI Ce 213	18.1	17.9	- .2	WI Cg 26	49.1	46.1	-3.0
WI Ce 232	33.0	32.5	- .5	WI Df 47	40.7	37.4	-3.3
WI Ce 233	6.5	2.6	-3.9	WI Df 51	42.7	31.3	-11.4
WI Ce 242	18.4	17.9	-.5	WI Df 60	32.2	30.1	-2.1

Table 6. - Hydrologic parameters used in the calibrated steady-state flow model

[in/yr, inch per year; ft/d, feet per day]

Model parameter	Value used in the calibrated ground-water-flow model
Recharge	14 in/yr
River-bottom conductivity	3.75 ft/d
Horizontal hydraulic conductivity of the Salisbury aquifer (layers 1 and 2)	layer 1: 204 ft/d layer 2: 313 ft/d
Horizontal hydraulic conductivity of the confining bed (layer 3)	0.01 ft/d
Vertical hydraulic conductivity of the confining bed (layer 3)	0.001 ft/d
Horizontal hydraulic conductivity of the paleochannel sediments (layers 3 and 4)	450 ft/d
Horizontal hydraulic conductivity of the Manokin aquifer (layers 4 and 5)	54 ft/d

Table 7. - Comparison of measured and simulated heads in the Manokin aquifer

Well number	Date measured	Head, in feet, related to sea level	Simulated head, in feet, related to sea level	Simulated head minus measured head, in feet
WI Ce 212	11-17-67	17.3	17.9	0.6
WI Cf 78	11-29-62	21.9	23.2	1.3

EXPLANATION

— 20 — CONTOUR OF THE SIMULATED POTENTIOMETRIC SURFACE. NUMBER IS THE ALTITUDE OF HEAD, IN FEET ABOVE SEA LEVEL. CONTOUR INTERVAL IS 5 FEET.

24 / 1/64 WELL CONTROL POINT. TOP NUMBER IS THE MEASURED WATER LEVEL, IN FEET ABOVE SEA LEVEL. BOTTOM NUMBER IS THE MONTH AND YEAR WATER LEVEL WAS MEASURED.

Figure 27. - Simulated potentiometric surface of the Manokin aquifer and measured water level in three wells.

Table 8. - Estimated and simulated baseflow in streams within the model area

Location	Estimated annual average baseflow (assumes 14 inches recharge)[1] (cubic feet per second)	Simulated mean base flow (cubic feet per second)	Basin area (square miles)
Leonard Pond Run, at Leonard Pond	14	9.1	13.4
Connelly Mill Branch	3.7	1.2	3.66
Little Burnt Branch, at Jersey Road	3.5	0.48	3.39
North Prong Wicomico River, at Naylor Mill Road	26	24	24.8
Brewington Branch	3.9	0.82	3.89
Middle Neck Branch	5.7	9.6	5.6
North Prong Wicomico River, at Isabella Street Bridge	38	44	37.1
Beaglin Branch	2.1	0.020	2.1
Beaverdam Creek at Schumaker Pond	19	16	19.5

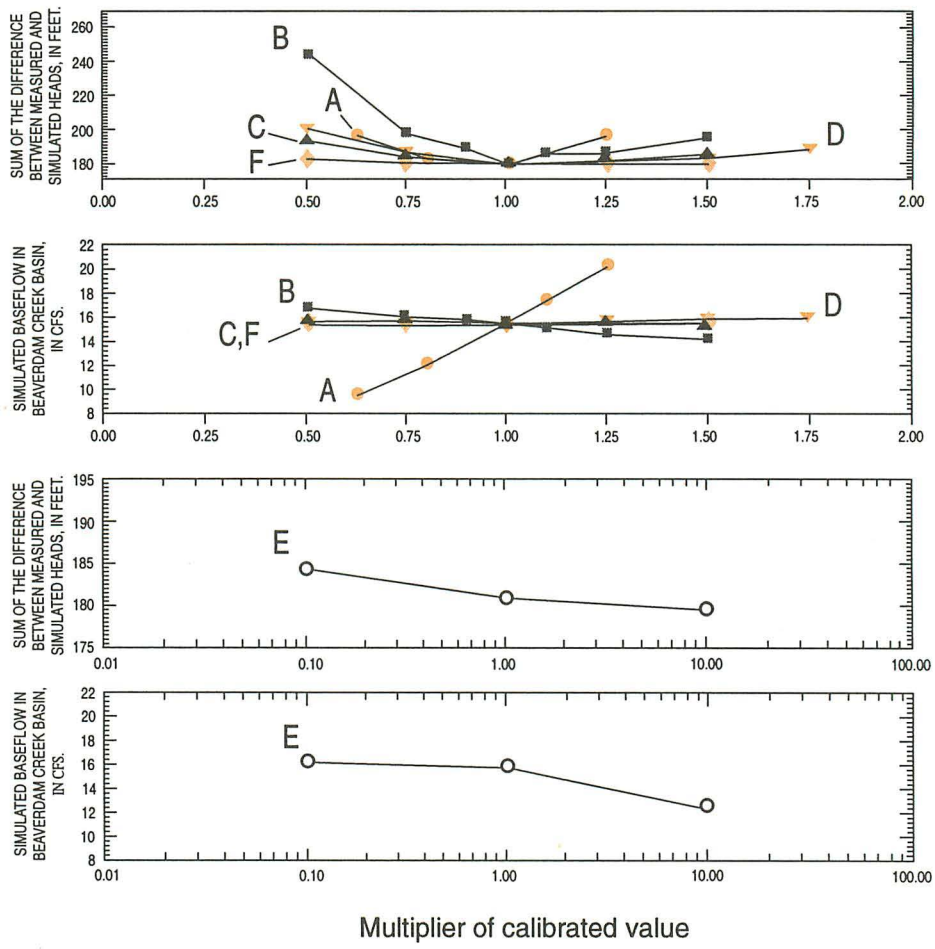
[1] Estimated by multiplying the drainage area by 14 inches per year recharge value.

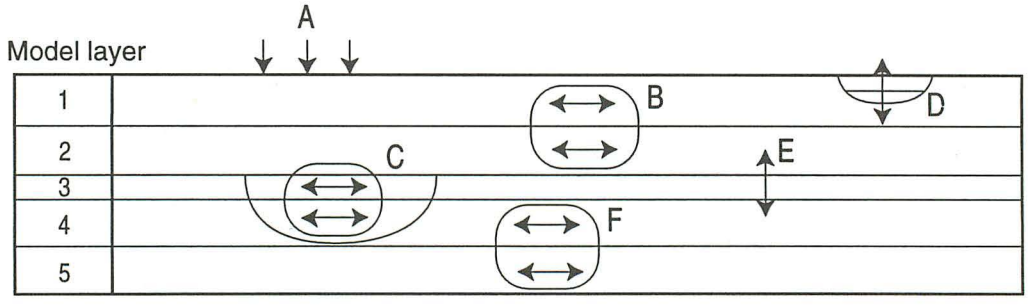
Possible explanations for the difference between simulated and estimated baseflow include, (1) difference between modeled sub-basin area and actual sub-basin area, (2) 14 in/yr of recharge is an inaccurate value, (3) lack of an exact match between simulated and measured heads, (4) simulated sub-basin flow divides are different from actual sub-basin divides, and (5) erroneous head differential between river stage and aquifer head caused by averaging of stage and head over the model-cell area. This last problem is most pronounced in large model cells where discretization of stream segments may not be accurate. Also, estimates of baseflow based on basin area and average annual recharge do not take into account diversion of water to wells.

The volumetric budget for the steady-state flow model under 1993 pumping conditions is given in table 9. The difference between total inflow and outflow is caused by rounding. The ground-water budget calculated by the model indicates that recharge from precipitation accounts for about 85 percent of total inflow (50 Mgal/d) with the remaining inflow coming from constant-head cells (10 percent or 5.9 Mgal/d) and river leakage (5 percent or 2.8 Mgal/d). Total outflow was divided between river leakage (77 percent or 45 Mgal/d), flow to the tidal part of the Wicomico River (8 percent or 4.8 Mgal/d), and production wells (15 percent or 9 Mgal/d). River leakage into the model is a result of leakage from ponded streams where the modeled stage in the pond is greater than the aquifer head. River inflow also occurs in the model in some upstream river reaches. This is an erroneous situation resulting from averaging in the model of aquifer head and river stage in the larger model cells. The volume of water entering the model in these cells is not great enough to adversely affect the model performance.

SENSITIVITY ANALYSIS OF STEADY-STATE MODEL

Various model parameters were varied independently across a range of values considered reasonable based on available field data. The effects of the changes on the calibrated model were observed and plotted with other variables (fig. 28). The model output parameters used in the sensitivity analysis included the sum of the absolute differences between the simulated heads and measured heads and simulated baseflow in Beaverdam Creek. Model input parameters considered in the analysis include recharge, horizontal hydraulic conductivity of aquifers, river-bottom hydraulic conductivity, and vertical leakance across the confining bed (layer 3). The results of the analysis are shown in figure 28.





A: Recharge
B: Horizontal hydraulic conductivity in Salisbury aquifer
C: Horizontal hydraulic conductivity in paleochannel
D: River bottom hydraulic conductivity
E: Vertical leakance across confining bed
F: Horizontal hydraulic conductivity in Manokin aquifer

Figure 28. - Effects on simulated head match and baseflow by varying model parameters.

Table 9. - Steady-state volumetric budget under 1993 pumping conditions

[ft^3/d, cubic feet per day; Mgal/d, million gallons per day]

Inflow and outflow sources	Inflow			Outflow[1]		
	10^6 ft^3/d	Mgal/d	Percent	10^6 ft^3/d	Mgal/d	Percent
Constant head	0.79	5.9	10	0.64	4.8	8
Production wells	0	0	0	1.2	9	15
Recharge	6.7	50	85	0	0	0
River leakage	0.37	2.8	5	6.0	45	77
Total	7.9	59	100	7.8	59	100
Error (Inflow - Outflow)[2]	0.1	0	0			

[1] Evapotranspiration is not treated as a separate outflow; rather, it is accounted for in the recharge term.
[2] Differences between error given in cubic feet per day and million gallons per day are caused by rounding.

Model heads are particularly sensitive to a decrease in horizontal hydraulic conductivity in the Salisbury aquifer. Reducing this parameter by half increased the sum of absolute differences in heads by 64 ft. Baseflow is most sensitive to a change in recharge. Increasing the recharge by a factor of 1.25 (or to 17.5 in/yr) increased baseflow to Beaverdam Creek by 4.7 ft^3/s. Decreasing recharge by a factor of 0.625 (or to 8.75 in/yr) lowered baseflow by 6.1 ft^3/s to Beaverdam Creek.

An order-of-magnitude variation in the vertical leakance across the confining bed has little apparent impact on simulated water levels, in part because there were only three head measurements available in the underlying Manokin aquifer. Simulated baseflow to Beaverdam Creek, however, decreased with an increase in vertical leakance. The vertical hydraulic gradient in the majority of the Beaverdam Creek Basin is downward from the Salisbury aquifer to the Manokin aquifer and a simulated increase in leakance caused more water to move downward from the Salisbury aquifer, thus reducing the amount of water discharged to streams.

ESTIMATION OF CONTRIBUTING AREAS

Water entering the well screens in the Salisbury aquifer within the study area is derived from different zones. Two important terms that describe these zones are (1) contributing area, and (2) zone of transport (U.S. Environmental Protection Agency, 1987). The contributing area of a well is defined as the land surface area (recharge area) through which infiltrating water reaches the well. The zone of transport is defined as the projected area surrounding a pumping well, bounded by lines of equal times, through which a water particle may travel and reach the well. The zone of contribution is a special case of the zone of transport where the water particles end at the water table.

Contributing areas were determined using output from the calibrated, steady-state ground-water-flow model and the U.S. Geological Survey particle-tracking program MODPATH (Pollock, 1989). MODPATH calculates ground-water velocity based on intercell flow rates and heads obtained from MODFLOW, and estimated porosity of aquifers and confining beds. Particles placed on the faces of model cells representing pumping wells were back-tracked along ground-water-flow lines until they reached the water-table surface. Particles were stopped at 1, 10, and 20 years to determine the 1-year, 10-year, and 20-year contributing areas. For instance, the 1-year contributing area represents water that reaches a well within 1 year of entering the ground-water system. The contributing areas are shown in figures in this part of the report and on plate 2 at the end of the report. On plate 2, contributing areas for the Park well field are based on simulations of the 1993 pumping alternatives, which resulted in the largest contributing areas.

In addition to the output from the ground-water-flow model output (heads and budget terms), MODPATH requires the input of porosity arrays to calculate particle velocity. Porosity of the aquifers and confining bed were estimated based on the general sediment type (clay, silt, sand, or gravel) (Freeze and Cherry, 1979). This information was necessary for the calculation of flow velocities made by the particle-tracking routine. The porosities of layers 1 and 2 were set at 30 and 25 percent, respectively. These values are typical for medium- to coarse-grained sands. The porosity of layer 3 varies from 55 percent in the areas representing the confining bed (clay) to 25 percent in the paleochannel area. The porosity in layer 4 varies from 30 percent in areas representing the Manokin aquifer to 25 percent within the paleochannel. Layer 5, representing the bottom half of the Manokin aquifer, was assigned a porosity of 30 percent.

Contributing areas are dependent on the complex behavior of ground-water-flow paths and the hydrogeologic framework. Aquifer materials are heterogeneous and hidden from direct observation, which prevents the development of an exact conceptual picture of the system. It is important to understand that variations in contributing areas are dependent on conceptualization of the aquifer system (Riley and Pollock, 1989). Because the particle-tracking calculations were based on steady-state conditions, the velocity calculations represent average conditions. In reality, particle velocities change continually in response to variations in water levels caused by cyclical changes such as recharge, and non-cyclical changes, such as pumpage. In the Salisbury aquifer, over an extended period, however, changes in velocity tend to fluctuate about an average, and results from the steady-state and transient simulations will be similar. Reilly and Pollock (1995) determined that if the ratio of the mean travel time to the period of the cyclic stress (such as seasonal changes in recharge) was much greater than one, then the transient area contributing recharge to wells was similar to the area calculated using an average steady-state condition. This is true only if the effect of non-cyclical stresses on the system is minimal. This is the case in the Salisbury aquifer within the study area. The cyclical changes in recharge are on the order of 1/2 year (seasonal) to several years. The mean travel times for particles tracked from the Paleochannel and Park wells to their recharge areas, however, were generally greater than 50 years. Also, water levels in the Salisbury aquifer are only minimally affected by changes in pumpage (fig. 10)

The following discussion describes the contributing areas resulting from particles tracked backwards from pumping wells in the Paleochannel and Park well fields under 1993 pumping conditions and under estimated 2010 pumping conditions. The effect on contributing areas from model uncertainty, changes in aquifer porosity, and drought are illustrated using well 1 (WI Ce 200) in the Paleochannel well field. These simulations are meant to show how variations in the flow model affects the size and shape of the contributing areas in general.

PALEOCHANNEL WELL FIELD

Contributing Areas Based on Pumping Conditions in 1993

Contributing areas of Paleochannel wells 1 and 2 (WI Ce 200 and 241, respectively), representing ranges in travel times of 0 to 1, 1 to 10, and 10 to 20 years, were simulated based on 1993 pumping conditions (figs. 29 and 30). Contributing areas representing a range in travel time of 20 to 50 years, along with the previous contributing areas, are shown on plate 2. Fifty particles were placed on each face of model cells representing well-screen positions, and were tracked backwards to the water-table surface (top of model layer 1). The final particle locations form irregular areas that represent contributing areas of the simulated wells.

In the simulation in which well 1 is pumped (fig. 29), two major contributing areas result; one extends outward from the well approximately 2 mi, and a second is disconnected from the well and is approximately 1 mi long. The large gap between the well and the contributing area north of Log Cabin Road is a result of water discharging to Little Burnt Branch. Water entering the saturated zone between the well and the contributing area is captured by Little Burnt Branch before reaching the well. Most of the contributing area lies within an agricultural area.

The top part of the well screen for well 1 (model layer 3) intercepts most of the water within the 20-year period. Only two relatively small areas within the 10- to 20-year contributing area supply water to the deeper part of the well screen (model layer 4) in the more productive part of the paleochannel. The shortest travel time is approximately 120 days and occurs along a vertical flow path from the water table adjacent to well 1.

The contributing area representing a travel time range of 20 to 50 years extends back to the modeled basin boundary (pl. 2). The angular shapes of the contributing area on the west and north sides are caused by the coarse discretization along the basin boundary in those areas. A finer grid along the basin boundary would probably result in an extension of the contributing area back to basin boundary.

In the simulation in which well 2 is pumped, the contributing area extends eastward from the well approximately 2.2 mi (fig. 30). Most of the water reaching the well within 20 years enters the top part of the well screen (layer 3). Water entering the middle section of the well screen (layer 4) is recharged mostly from the east side of U.S. Route 13. No water reaches the bottom part of the well screen (layer 5) within 20 years. The minimum travel time is approximately 160 days.

Ground water entering the wells in the Paleochannel well field comes primarily from the Salisbury aquifer; however, the Manokin aquifer supplies some water to the deeper parts of the well screens. Cross sections along model rows extending through wells 1 and 2 show that particles flow

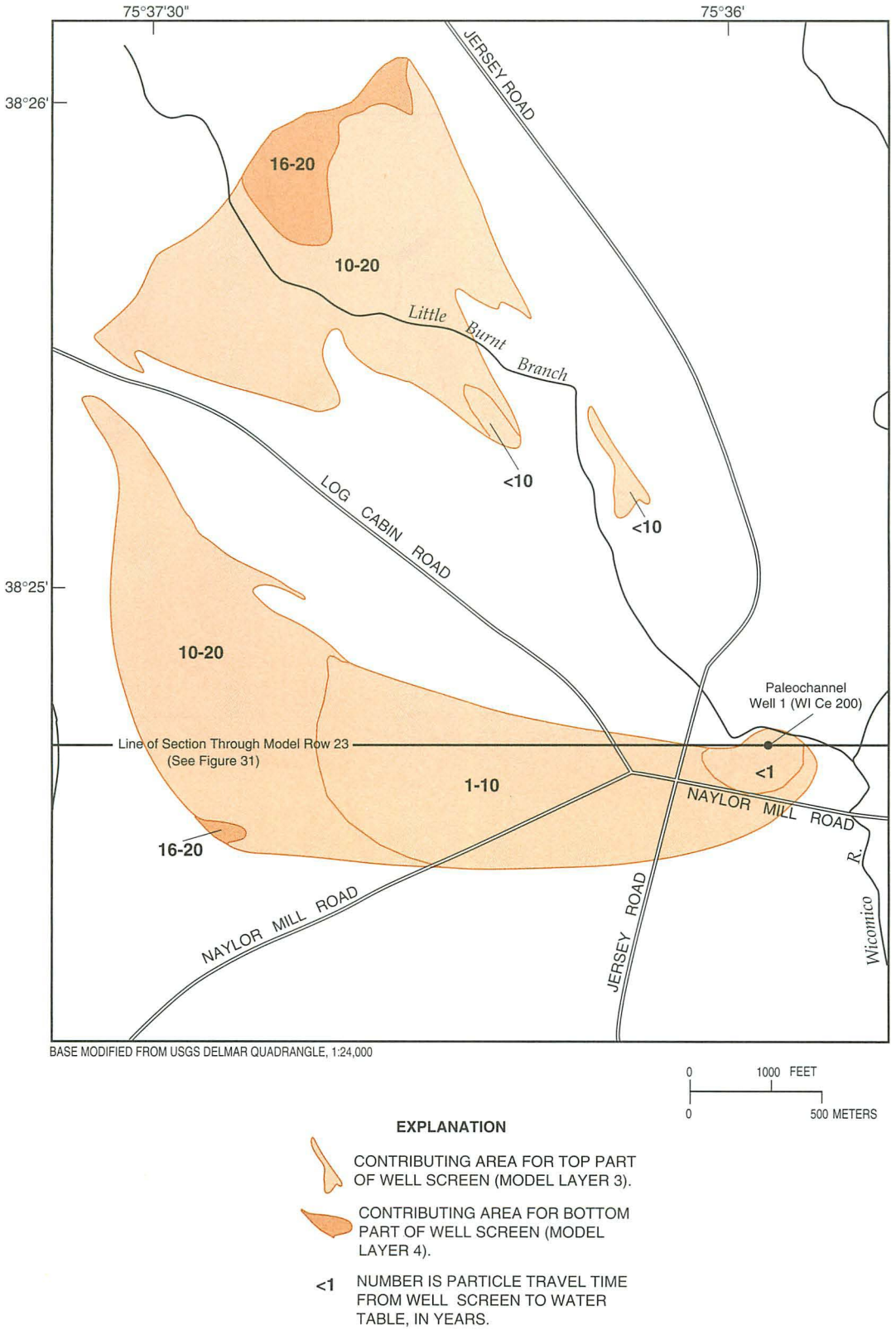


Figure 29. - Contributing areas for Paleochannel well 1 based on 1993 pumping conditions.

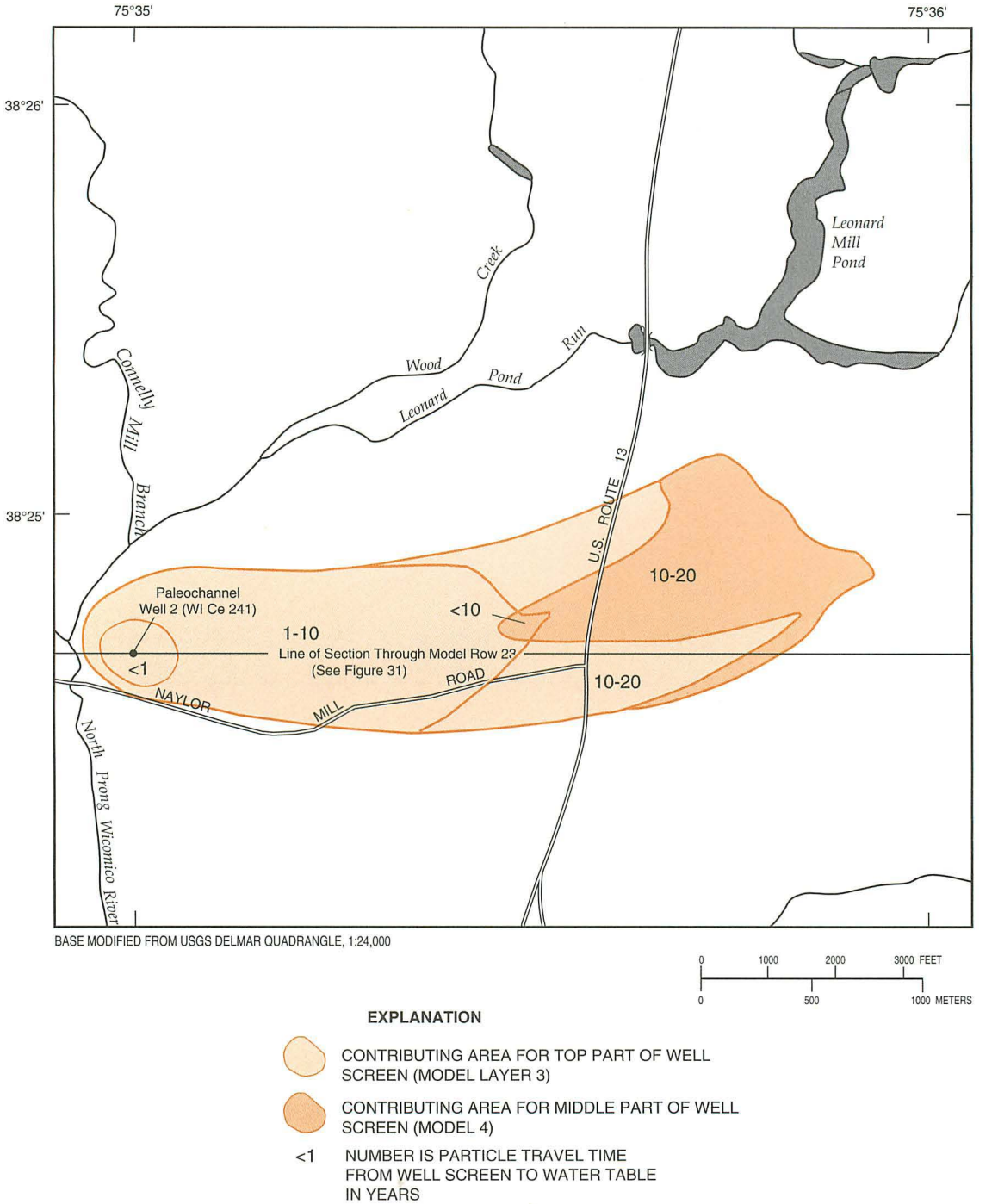


Figure 30. - Contributing areas for Paleochannel well 2 based on 1993 pumping conditions.

toward the wells from both the Salisbury (layers 1, 2 and 3) and Manokin (layers 4 and 5) aquifers (fig. 31). For clarity, only a few selected particle paths are shown in the cross sections, because particle paths were projected onto one model row. Intersecting and foreshortened particle paths are a result of this projection. The particles were stopped after back-tracking 20 years. Particles entering the west side of the cross section showing well 2 appear to reverse gradient within a short distance. This is a visual effect created by projecting particle paths onto a single plane (row 27). In reality, these particles move downgradient within the Manokin aquifer from the west-northwest and approach the surface where the Manokin is breached by the paleochannel. As the particles approach well 2, they are captured by the well.

Variations in Contributing Areas Based on Parameter Uncertainty

The accuracy of the contributing areas drawn around supply wells is linked directly to the uncertainty of parameter values used in the ground-water-flow model. To illustrate how the contributing areas are affected by changes to model parameters adjusted during model calibration, a backward-track simulation was made from Paleochannel well 1 based on a model simulation using a 7-percent increase (0.98 in/yr) in recharge over the value used in the calibrated model (14 in/yr). This change resulted in a slightly different match of simulated and measured heads--sum of the absolute difference between simulated and measured heads of 185 ft as compared to the calibrated value of 181 ft--and the same simulated baseflow to Beaverdam Creek (15.7 ft^3/s). The 10-year contributing area based on this increase in recharge resulted in a somewhat different pattern than that produced using the calibrated value of 14 in/yr. The large west-trending portion of the contributing area narrowed and shifted northward slightly. The northeast-trending portion is approximately 500 ft longer than the contributing area based on the calibrated model.

Variations in Contributing Areas Caused by Changes in Porosity

The particle-tracking routine uses porosity values in concert with head output from the ground-water-flow model to calculate particle velocities. Porosity arrays assigned to each model layer were chosen based on the general type of sediment (gravel, sand, clay, etc.). The values are estimates representing average porosity, and, as such, are imprecise. Because porosity is necessary for calculating velocities used in the determination of contributing areas, it is important to evaluate the effect that uncertainty in this parameter has on the results. The 10-year contributing area of production well 1 (WI Ce 200), calculated using estimated porosity described earlier in the report, was compared to the 10-year contributing areas calculated using a 10-percent increase and 10-percent decrease in porosity. The results indicate that a 10-percent decrease in porosity expands the contributing area approximately 500 ft; the decrease in porosity increases particle velocities and allows more particles to reach the well within the same time span. A 10-percent increase in porosity shrinks the contributing area by approximately 500 ft; the increase in porosity decreases particle velocities and results in fewer particles reaching the well within the same time span. If in error, the calibrated model probably underestimates aquifer porosity.

Contributing Areas Based on a Simulated Drought

A drought represented by a 40-percent decrease in recharge (8.4 in/yr) was simulated (steady state) to determine the effect of such a condition on the contributing area of Paleochannel well 1. To simulate the drought, the river cells in the model were converted to drain cells. This was necessary because decreasing the recharge in the model resulted in aquifer heads below stream levels in many locations, which induced an unrealistic amount of inflow from the streams. Converting river cells to drains corrects this problem by preventing water from flowing into the model when the aquifer heads drop below the drain altitude. This is an extremely simplified simulation of the effects of a drought condition in that it does not take into account the changing rates of evapotranspiration and the increase in ground-water withdrawals typically associated with long periods of reduced precipitation. Also, a steady-state simulation greatly exaggerates the effect of a drought on the hydrologic system.

The steady-state simulation of a drought condition results in heads somewhat lower than would occur over a typical drought period of approximately 3 years (fig. 32). The steady-state simulation calculates heads that have reached equilibrium with the reduced recharge rate--a condition not met within a 3-year period. A transient model simulation was run to compare the steady-state heads with head conditions during 3 years of drought. Three stress periods, 1 year in duration, were simulated using a recharge rate of 8.4 in/yr and the 1993 pumpage distribution. The transient model used the same hydraulic parameters as the steady-state model, with the addition of specific yield and storage coefficient. An estimated specific yield of 0.1 was used for layer 1. This value is within the range for unconfined aquifers (Freeze and Cherry, 1979). An estimated storage coefficient of 0.0001 was used for layers 2, 3 (within the paleochannel), 4, and 5. This value is within the range for confined aquifers. The storativity of the confining bed (layer 3) was set at $1.24\ 10^{-5}$. This number is a median value calculated from a series of constant-flow and consolidation tests performed on the confining bed near the Park well field (Wolff, 1970).

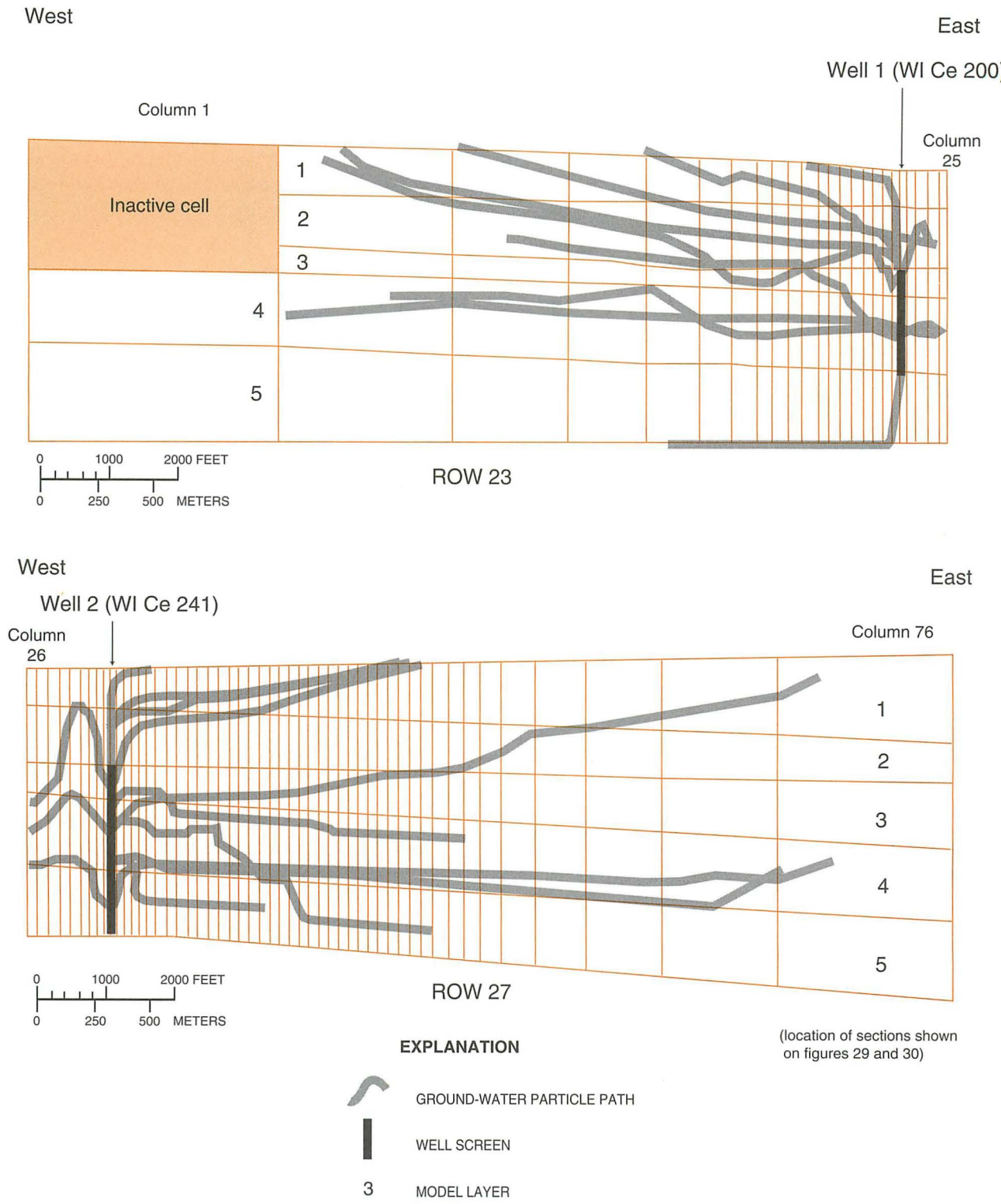


Figure 31. - Cross sections of particle paths back-tracked 20 years from Paleochannel wells 1 and 2 along model rows 23 and 27, respectively.

Heads simulated using the steady-state model were used as starting heads in the transient model.

Results from the transient simulation indicate that heads near the two well fields reach equilibrium within the 3-year simulation period. Heads upgradient from the well fields take longer to reach equilibrium. Therefore, the steady-state drought simulation results in heads lower than what might occur over just a 3-year period, thus overestimating the effects of the drought.

Backward-particle tracking based on the steady-state drought simulation indicates that the contributing area is broader than that determined by the non-drought simulation (figs. 29 and 33). The two branches of the non-drought contributing area have merged in the drought simulation.

Contributing Areas Based on Estimated Pumpage for 2010

The extent and shape of the contributing areas of pumping wells are related in part to the volume of water pumped. For example, the contributing area of a pumping well will generally increase in size with an increase in withdrawals from that well. However, an increase in withdrawals from nearby wells may capture some water previously flowing to the well, thereby moving the contributing area further afield.

Ground-water pumpage from the City of Salisbury well fields is estimated to increase to approximately 6 Mgal/d by the year 2010 (Resources Management Associates, Inc., 1977, as amended in 1993). This amount represents a 15-percent increase over 1993 pumping amounts. Assuming that the ratio of water supplied by the Paleochannel and Park well fields in 1993 will remain constant, the amount of water pumped from these well fields in 2010 will equal 4 and 2 Mgal/d, respectively. Contributing areas of wells 1 and 2 were calculated based on these pumpage figures (figs. 34 and 35, respectively).

Industrial, irrigation, and residential water withdrawals outside the city's service area were left at 1993 levels during the particle-tracking simulations. The number of industrial self-supplied users, the majority of which are canning and food processors, is not expected to increase significantly by 2010 (Resources Management Associates, Inc., 1977, as amended in 1993). Future pumpage from existing industrial users is difficult to determine because it is dependent on economic factors. Withdrawals for irrigation and self-supplied residential use will probably not increase significantly. The population density in the central part of Wicomico County outside of the City of Salisbury will probably increase gradually in the vicinity of small communities, with the majority of the land remaining in agricultural use, woodland, and swampland (Resource Management, Inc., 1977, as amended in 1993). The number of acres irrigated within the study area are not likely to increase because of the slow encroachment of residential housing outward from the City of Salisbury and because of the scarcity of large tracts of new land suitable for irrigation (Wayne Shaft, Agricultural Cooperative Extension Service, oral commun., 1995).

The southern section of the contributing area of Paleochannel well 1 increased about 200 ft on a side from the 1993 pumpage simulation. The northern section of the contributing area increased about 1,000 ft on a side from the 1993 pumpage simulation (fig. 34). The increase in contributing area in this location indicates that water is being diverted from baseflow to the upper reach of Little Burnt Branch.

The contributing area of Paleochannel well 2 extends approximately the same distance toward the northeast as under 1993 pumping conditions. The width of the contributing area, however, increased by as much as 600 ft over the area simulated under 1993 pumping conditions (fig. 35).

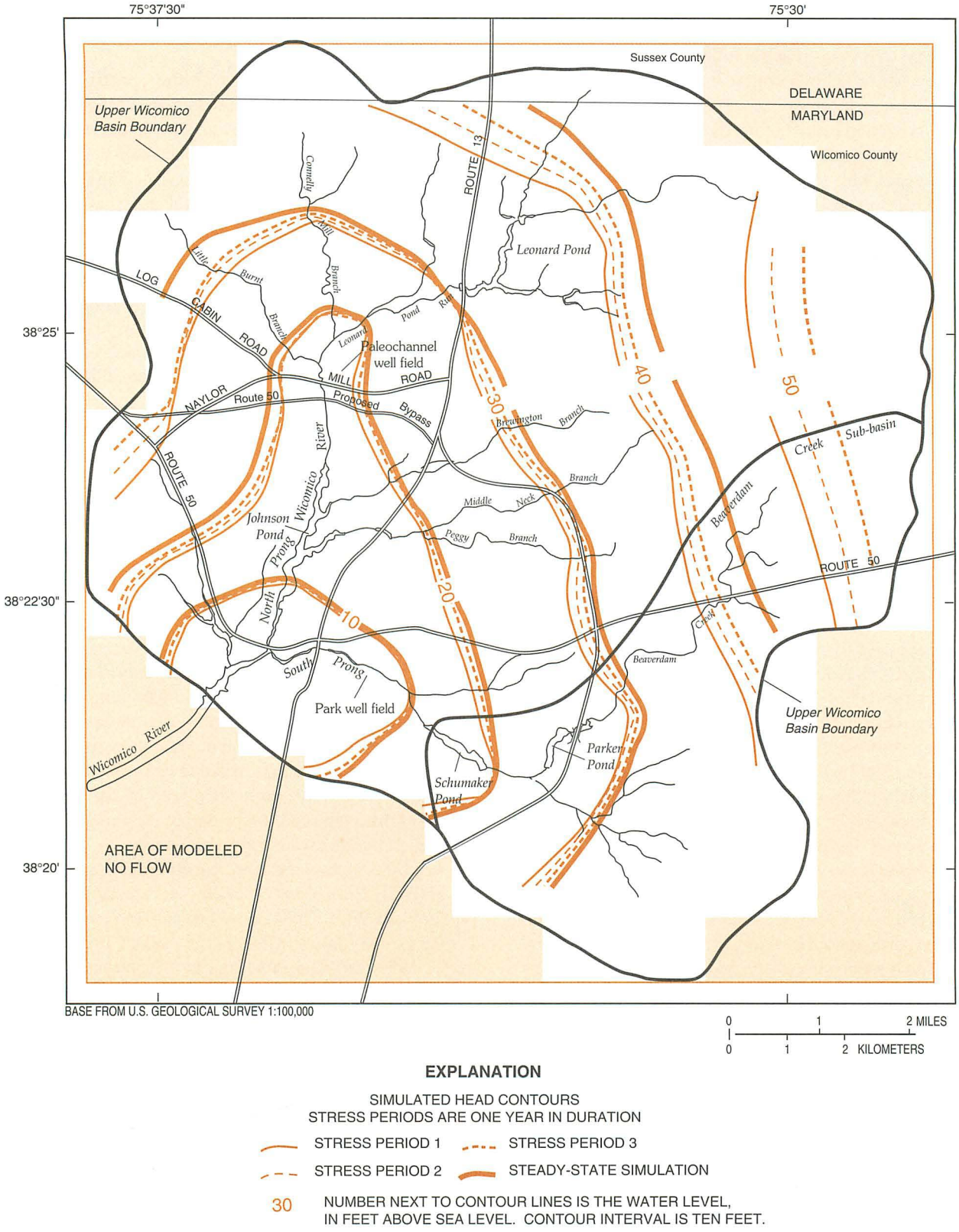
PARK WELL FIELD

Contributing Areas Based on Pumping Conditions in 1993

The average daily pumping rate for the Park well field during 1993 was 1.7 Mgal/d. Wells in the Park field are pumped at different times, in different combinations, and for varying durations. Since the steady-state ground-water-flow model can not reproduce the effects caused by changes in pumpage over time, the total 1993 pumpage value was divided evenly and assigned to each pumping well in the Park well field. The contributing areas resulting from this pumpage distribution is meant to represent an average condition. Twenty-five particles placed on the four sides and top of the model cells representing the pumping wells were used in the backward-track simulation.

Contributing areas representing ranges in travel times of 0 to 1, 1 to 10, and 10 to 20 years are shown in figure 36 for wells WI Ce 248 (2a), WI Ce 246 (6a), WI Ce 247 (7b), WI Ce 248 (8a), WI Cf 194 (10a), WI Cf 193 (14a), and WI Cf 195 (16a). Contributing areas representing a range in travel time of 20 to 50 years for these wells, along with the previous contributing areas, are shown on plate 2. The contributing areas extend approximately 10,000 ft upgradient from the pumping wells. Only three wells--wells 2a, 6a, and 8a--capture particles from both sides of Park Pond. The curve in the contributing areas reflects the configuration of the water-table surface, whereas the width is controlled by the amount of flow captured by the wells relative to the amount flowing to Park Pond. Water entering the ground-water system in the areas between the contributing areas discharges to Park Pond.

The only wells receiving recharge in less than 1 year are 2a, 6a, and 7b. The first particles to reach these wells arrived within 40, 270, and 300 days, respectively. The first



EXPLANATION

SIMULATED HEAD CONTOURS
STRESS PERIODS ARE ONE YEAR IN DURATION

STRESS PERIOD 1 — STRESS PERIOD 3

STRESS PERIOD 2 — STEADY-STATE SIMULATION

30 NUMBER NEXT TO CONTOUR LINES IS THE WATER LEVEL, IN FEET ABOVE SEA LEVEL. CONTOUR INTERVAL IS TEN FEET.

Figure 32. - Simulated Salisbury aquifer steady-state water table based on a drought condition compared to 1, 2, and 3 years of transient simulation under drought conditions.

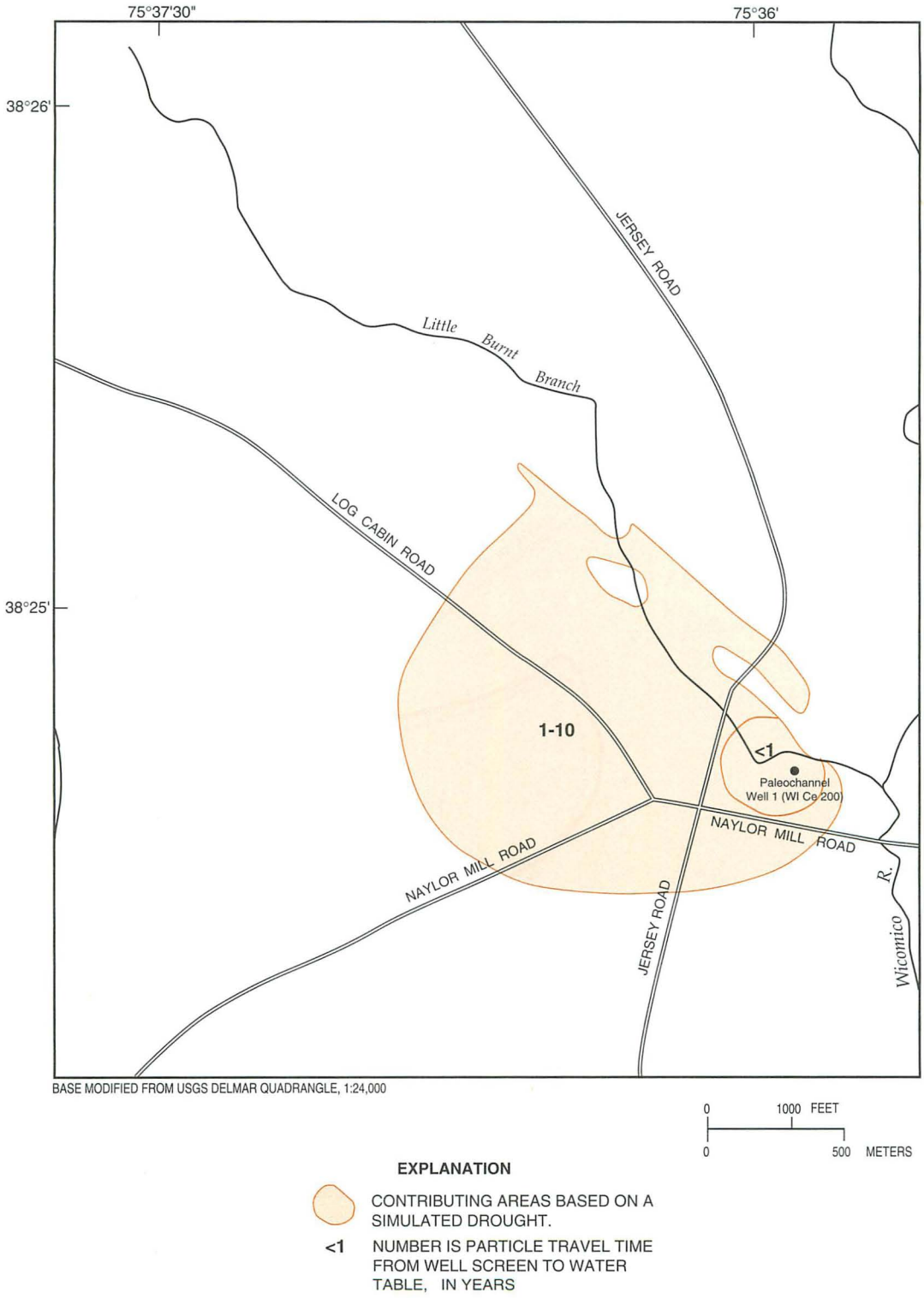


Figure 33. - Contributing areas for the Paleochannel well 1 based on a steady-state simulation of a drought condition.

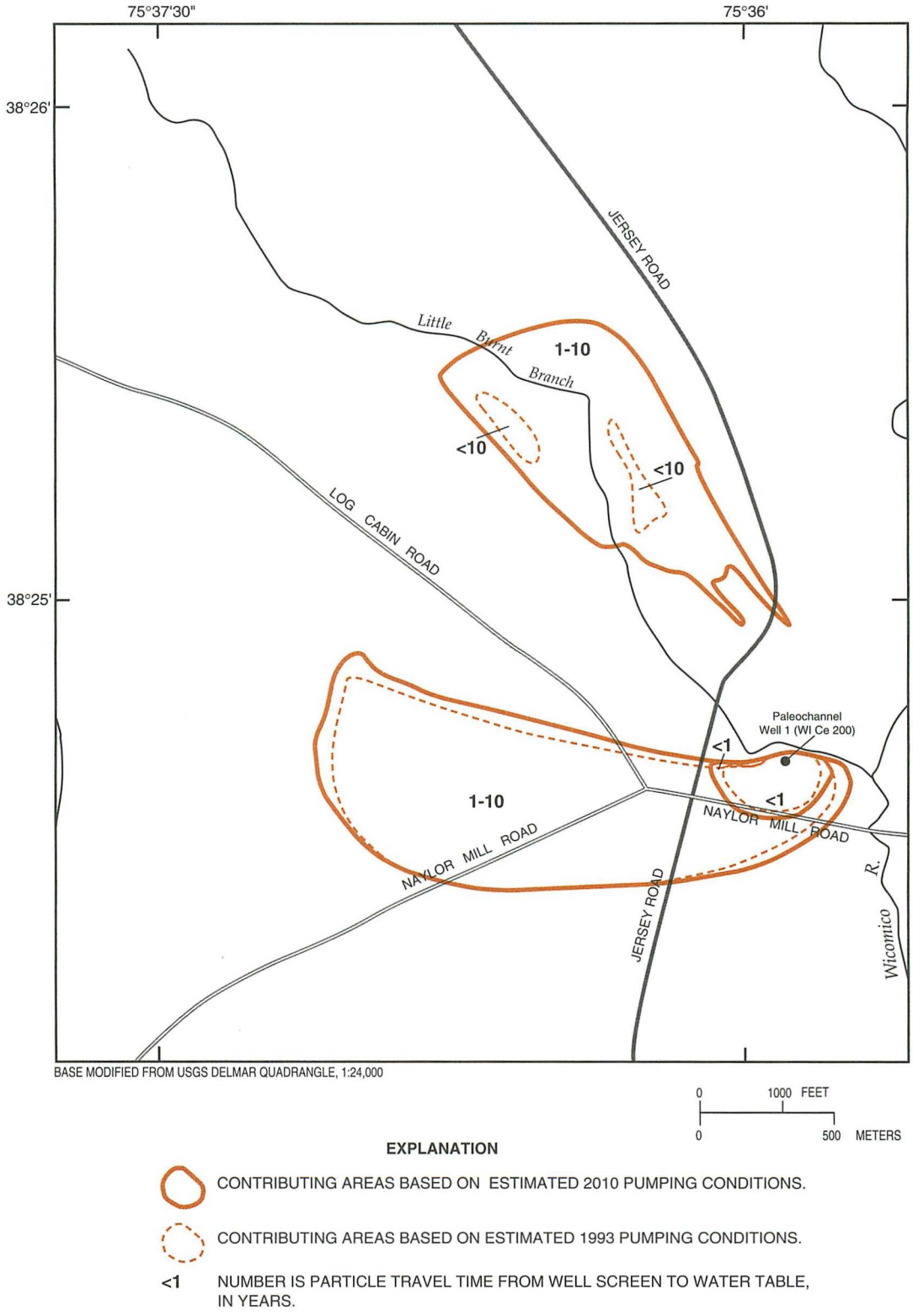


Figure 34. - Comparison of 1-year and 10-year contributing areas for Paleochannel well 1 under 1993 and estimated 2010 pumping conditions.

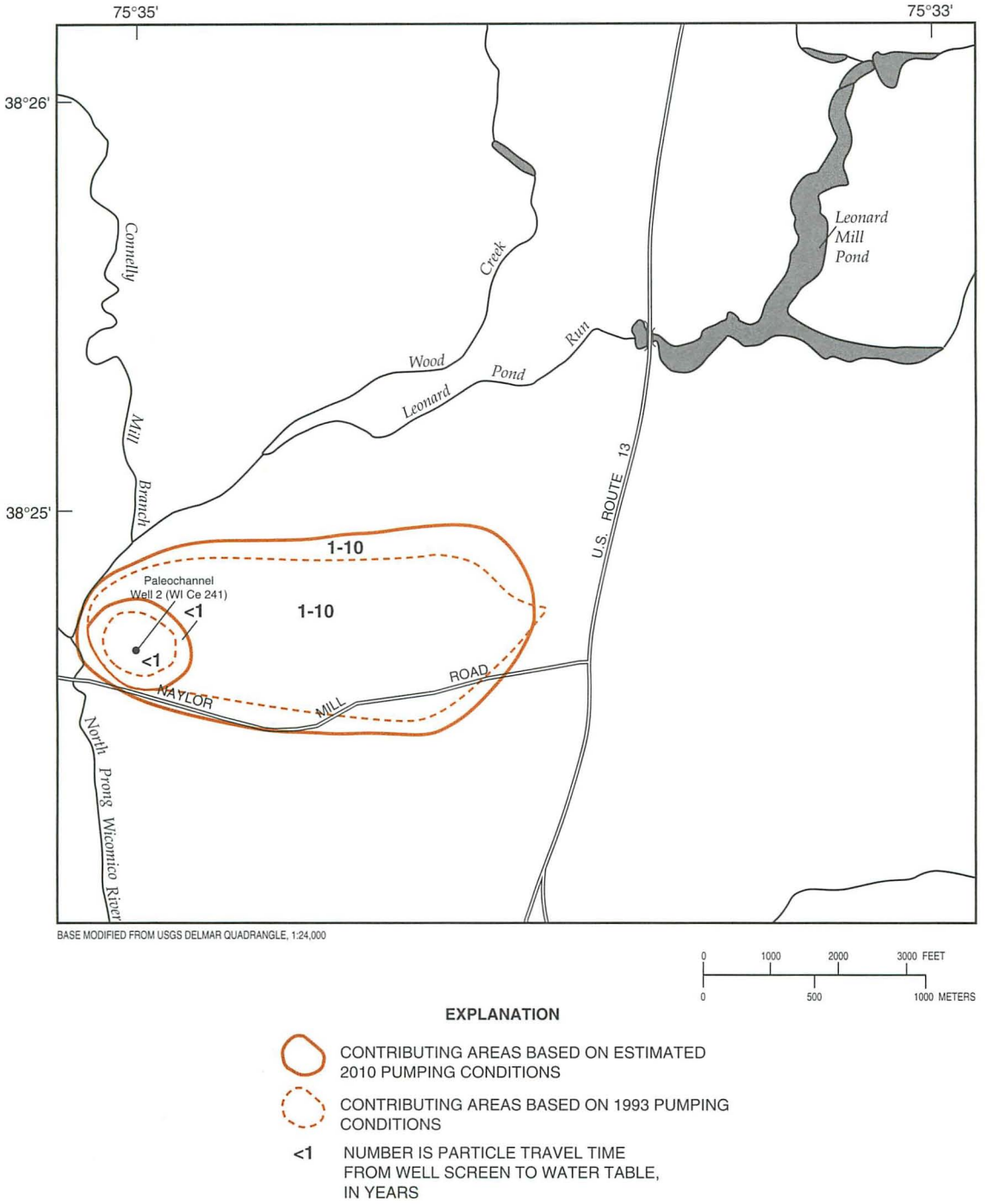
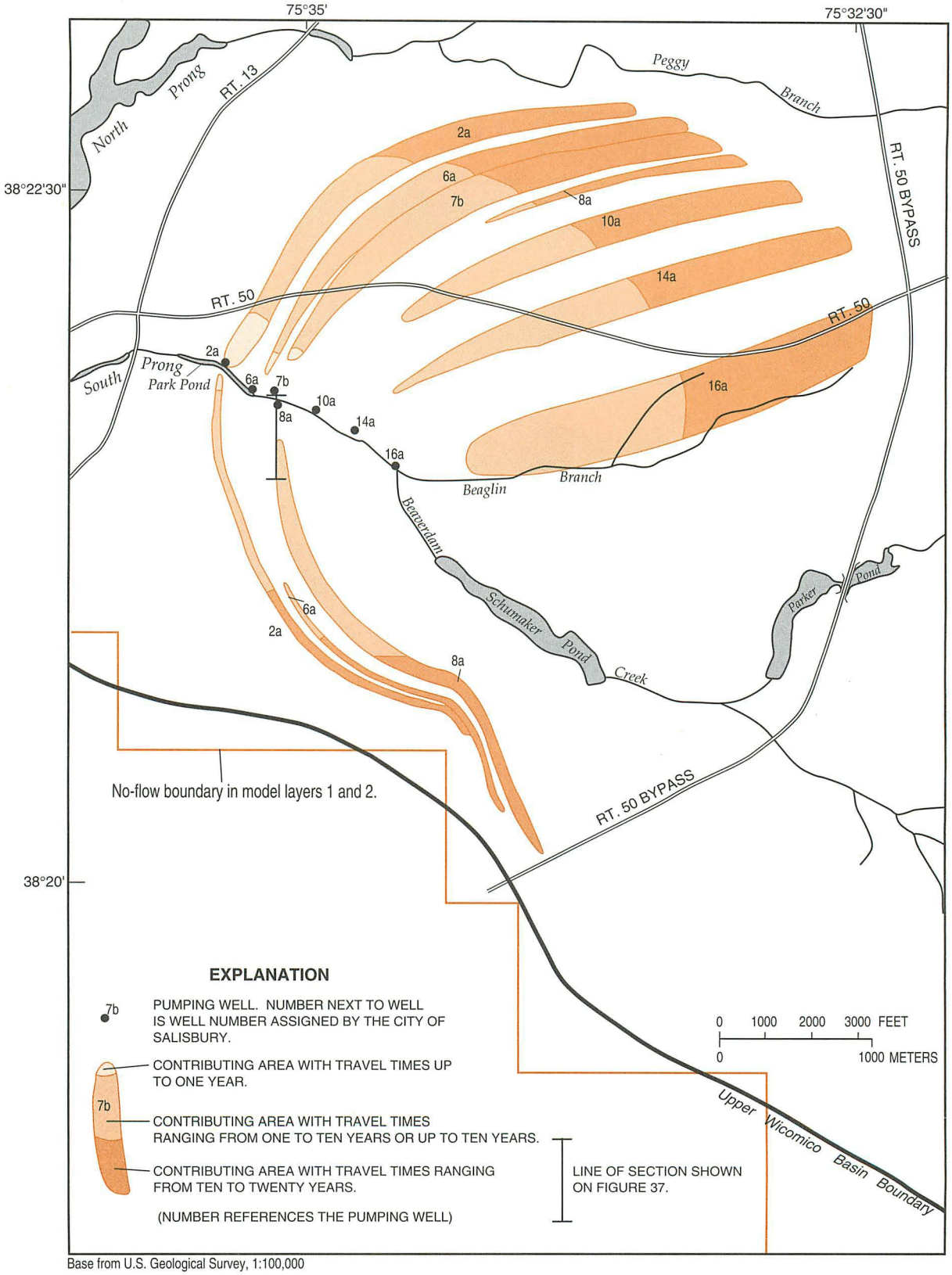


Figure 35. - Comparison of 1-year and 10-year contributing areas for Paleochannel well 2 under 1993 and estimated 2010 pumping conditions.



Base from U.S. Geological Survey, 1:100,000

Figure 36. - Contributing areas for Park wells 2a, 6a, 7b, 8a, 10a, 14a, and 16a based on 1993 pumping conditions.

particles to reach wells 8a, 10a, 14a, and 16a arrived within 500, 1,200, 400, and 460 days, respectively.

Under non-pumping conditions, ground water in the vicinity of Park Pond discharges to the pond. As discussed previously, there is probably a direct interconnection between the pond and aquifer. If pumping is great enough, water may flow out of the pond or baseflow to the pond may be reduced. Because of the level of finite-difference discretization in the ground-water-flow model, this effect could not be simulated. Simulated drawdowns associated with the pumping wells were less than actual drawdowns, caused in part by the averaging of the pumping rate over the model cell area. Because of the inability of the model to simulate the flow patterns immediately surrounding the pumping wells and pond precisely, the area between the wells and the contributing areas should be included in the model-calculated contributing areas.

Particles entering the flow system between a pumping well and a contributing area to a pumping well travel along shallow flow paths and discharge to the pond. This behavior is illustrated in a cross section extending through Park Pond and well 8a (WI Cf 248) (fig. 37). The distances between the wells and their contributing areas are dependent on the relative strength of the modeled sinks representing pumping wells and modeled sinks representing the pond. That is to say, if flow to model cells representing the pond is greater than flow to nearby model cells representing pumping wells, then water entering the flow system near the pumping wells may bypass the pumping cells and discharge to the pond cells.

Two wells, WI Cf 201 (well 17) and WI Cf 202 (well 18), were added to the Park well field in 1994. Pumping these wells in addition to the existing wells would have an effect over time of reducing the amount of stress at the individual well sites, thereby reducing the size of the contributing areas. This is illustrated in figure 38, where 1993 pumpage (1.7 Mgal/d) is assigned evenly to all of the wells in the Park well field. In general, the upgradient extent of the contributing areas is similar to that obtained in the simulation with only the seven wells pumping (fig. 36). However, the distance from the wells to their contributing areas is reduced for all wells except 2a. The combined contributing areas for wells 16a, 17, and 18 is both longer and broader than that for well 16a alone.

Contributing Areas Based on Estimated Pumpage for 2010

The estimated pumpage from the Park well field in 2010, based on the projections made in the 1993 Wicomico County water and sewerage plan, is 2 Mgal/d. This represents a 15-

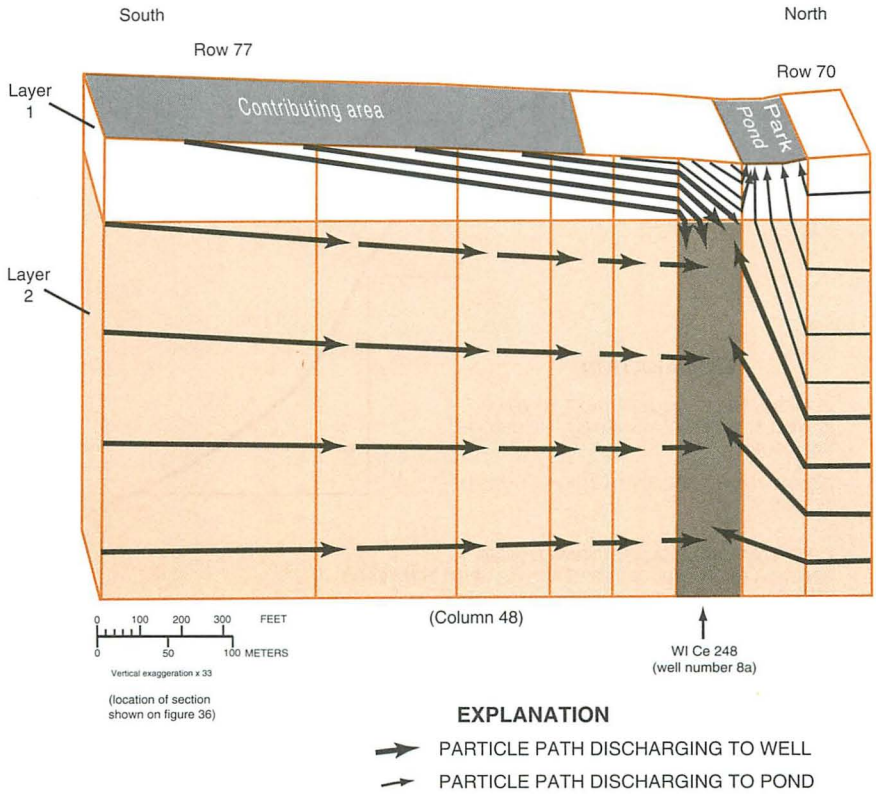
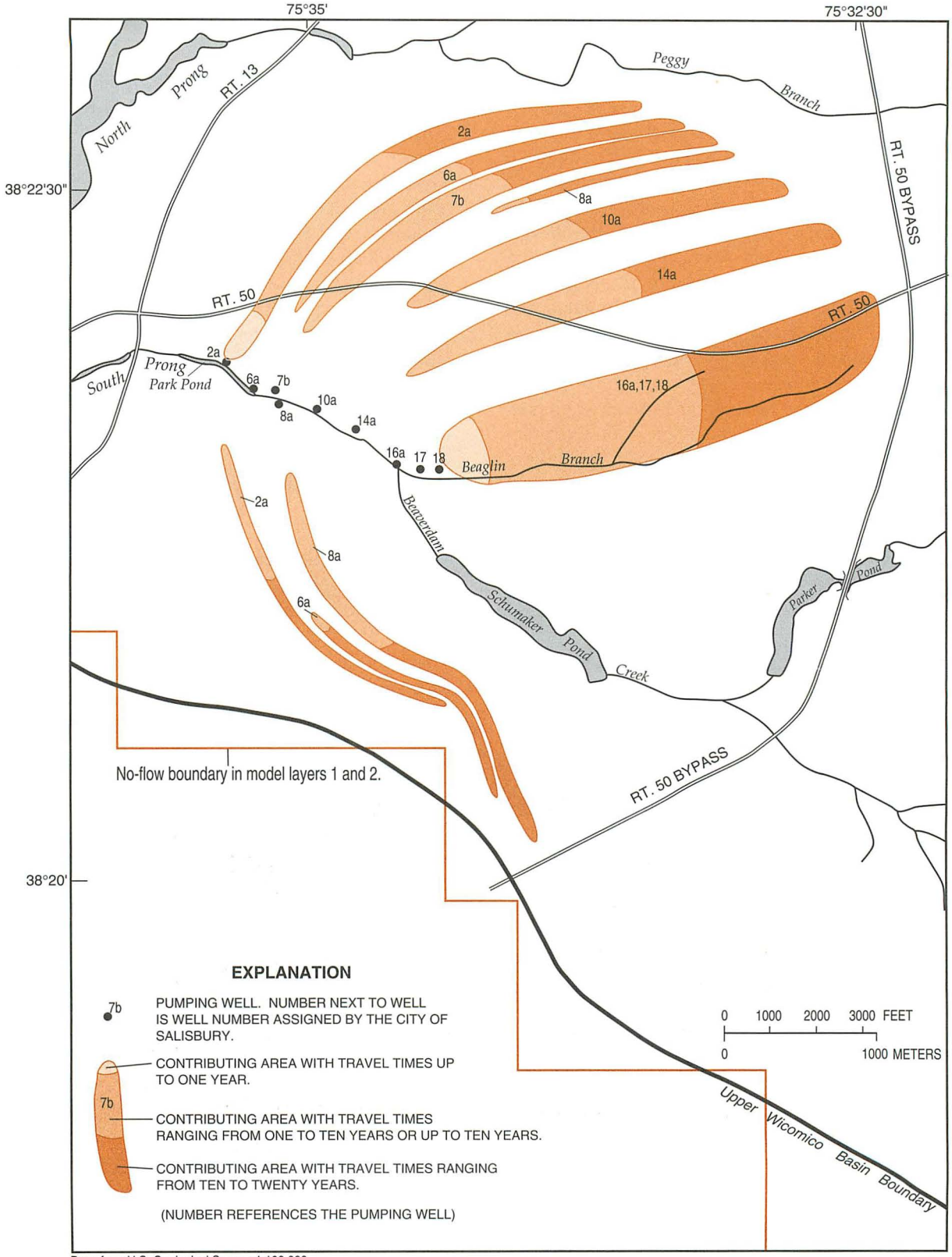


Figure 37. - Cross section along model column 48 showing the behavior of particle paths near Park well 8a and the Park Pond under 1993 pumping conditions.



Base from U.S. Geological Survey, 1:100,000

Figure 38. - Contributing areas for all Park wells based on 1993 pumping conditions.

percent increase over the 1993 level. Because the city added two production wells to the Park well field in 1993, increasing the number of wells to nine, the average simulated 2010 pumpage per well used in the model was less than that used in the 1993 simulation. This resulted in contributing areas smaller than those simulated under 1993 pumping conditions. To illustrate how the contributing areas change with a significant increase in pumpage, an arbitrary increase of 50 percent (or 3.4 Mgal/d) was used. Figure 39 shows the contributing areas based on a 50-percent increase in 1993 pumpage distributed evenly over the entire Park well field. Paleochannel well 1 was pumped at 4.2 Mgal/d and all other pumping wells were pumped at 1993 levels. Also shown on figure 39 are the contributing areas for wells 2a, 6a, 7b, 8a, 10a, 14a, and 16a under 1993 pumping conditions. The contributing areas based on the 50-percent increase in 1993 pumpage is as much as 1,000 ft on a side larger than the contributing areas based on the 1993 pumpage simulation.

FORWARD PARTICLE-TRACKING FROM POTENTIAL SOURCES OF CONTAMINATION BASED ON 1993 PUMPING CONDITIONS

To determine the possible flow direction from potential contaminant sites in the Upper Wicomico River Basin, particles were tracked forward from potential contaminant sites to discharge areas. Potential contaminant sites considered include CERCLA (Comprehensive Environmental Response, Compensation, and Liability Act) sites, a landfill, and sites with ground-water discharge permits. Also considered as potential contaminant sites, because of the possibility of spills (or future development along the right-of-way), are the proposed U.S. Route 50 Bypass, the Conrail railroad, and rivers and ponds near the well fields. Time increments were not assigned to the particle paths, because many factors, such as dispersion, attenuation of contaminants in the unsaturated zone and soil matrix, dilution, and chemical degradation, are not simulated. Particle tracking using MODPATH assumes conservative flow; that is, the hypothetical chemical constituent in the ground-water system would flow at the same rate as the ground water. In the forward-tracking simulations presented in this report, the particle paths are projected onto the top layer; some particle paths may pass through to deeper layers before exiting at boundaries such as the water table or pumping wells. Also, the simulations assume that the particles start at the water-table surface--unsaturated flow is not simulated in MODPATH. The simulations are based on the 1993 pumping levels used in the calibrated ground-water-flow model. In that simulation Paleochannel well 1 and Park wells 2a, 6a, 7b, 8a, 10a, 14a, and 16a were pumped.

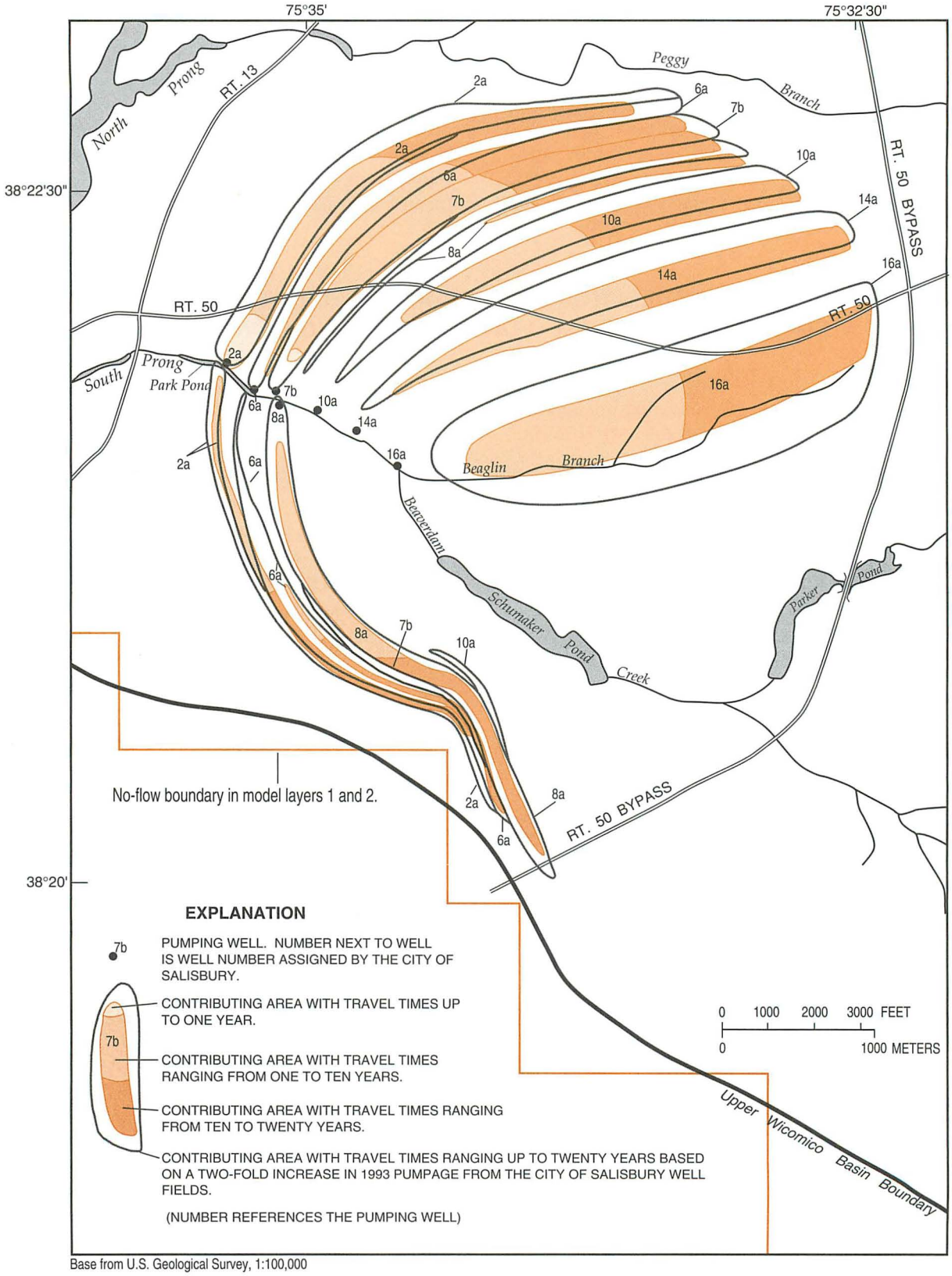
Particles tracked from the Newland Park landfill and the proposed U.S. Route 50 Salisbury Bypass are shown in figure 40. Twenty-eight particles were distributed over the top of the two model cells, which cover the existing Newland Park landfill and the proposed landfill expansion. The particles flowed in a southeasterly direction before discharging to river cells near Mitchell Pond. Particles entering the water table from the proposed U.S. Route 50 Salisbury Bypass flow south of the bypass and discharge to the North Prong Wicomico River. This section of road was given special consideration because it will pass within 0.3 mi of the Paleochannel well field.

Particles tracked from the Conrail railroad right-of-way enter the water table and flow toward the west (fig. 41). The particles discharge to the North Prong Wicomico River and its tributaries in the upper part of the basin and to the tidal part of the Wicomico River in the southern part of the basin. The particle paths are too shallow to reach the well screens in Paleochannel well 2 (WI Ce 241) under conditions when well 2 is not pumping. In a simulation in which well 2 was pumped instead of well 1, the minimum travel time between the railroad right-of-way and well 2 was approximately 2 years.

As previously discussed, the stretch of Little Burnt Branch that passes Paleochannel well 1 loses some of its flow during low baseflow conditions as a result of pumping from Paleochannel well 1. This condition is not likely to prevail during periods of high baseflow. Because the production well derives some of its flow from the stream during certain parts of the year, the stream becomes part of the contributing area and, therefore, a possible source of contamination. It seems unlikely, however, that the stream would constitute a serious risk of contamination to the well given that (1) flow from stream to well likely only occurs during several months out of the year, (2) a contaminant entering the stream would first be diluted in the stream, then in the aquifer, and (3) a contaminant in the stream would pass by the well field relatively quickly and only a small quantity of the contaminant would infiltrate into the ground-water system.

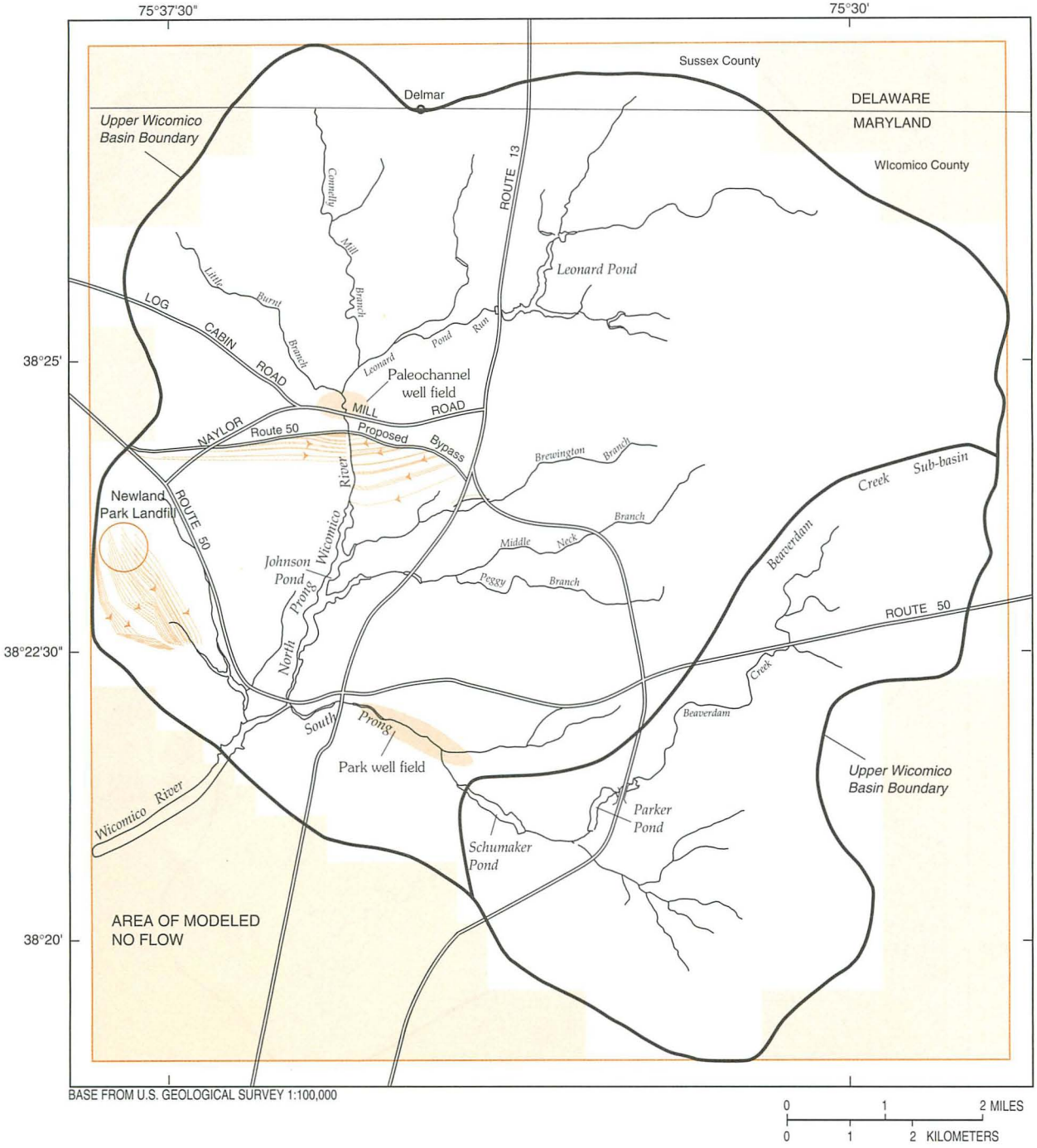
The steady-state ground-water-flow model simulated average seasonal conditions. In that simulation, the part of Little Burnt Branch that passes near Paleochannel well 1 was gaining water. Particles placed in those cells did not enter the ground-water system.

Wells in the Park well field may capture water from Park Pond during periods of low baseflow and high withdrawal rates. The pond area, therefore, should be considered part of the contributing area to the wells. As discussed earlier, the model could not adequately simulate the complex flow paths near the pond. The pond acted as a discharge area in model simulations.



Base from U.S. Geological Survey, 1:100,000

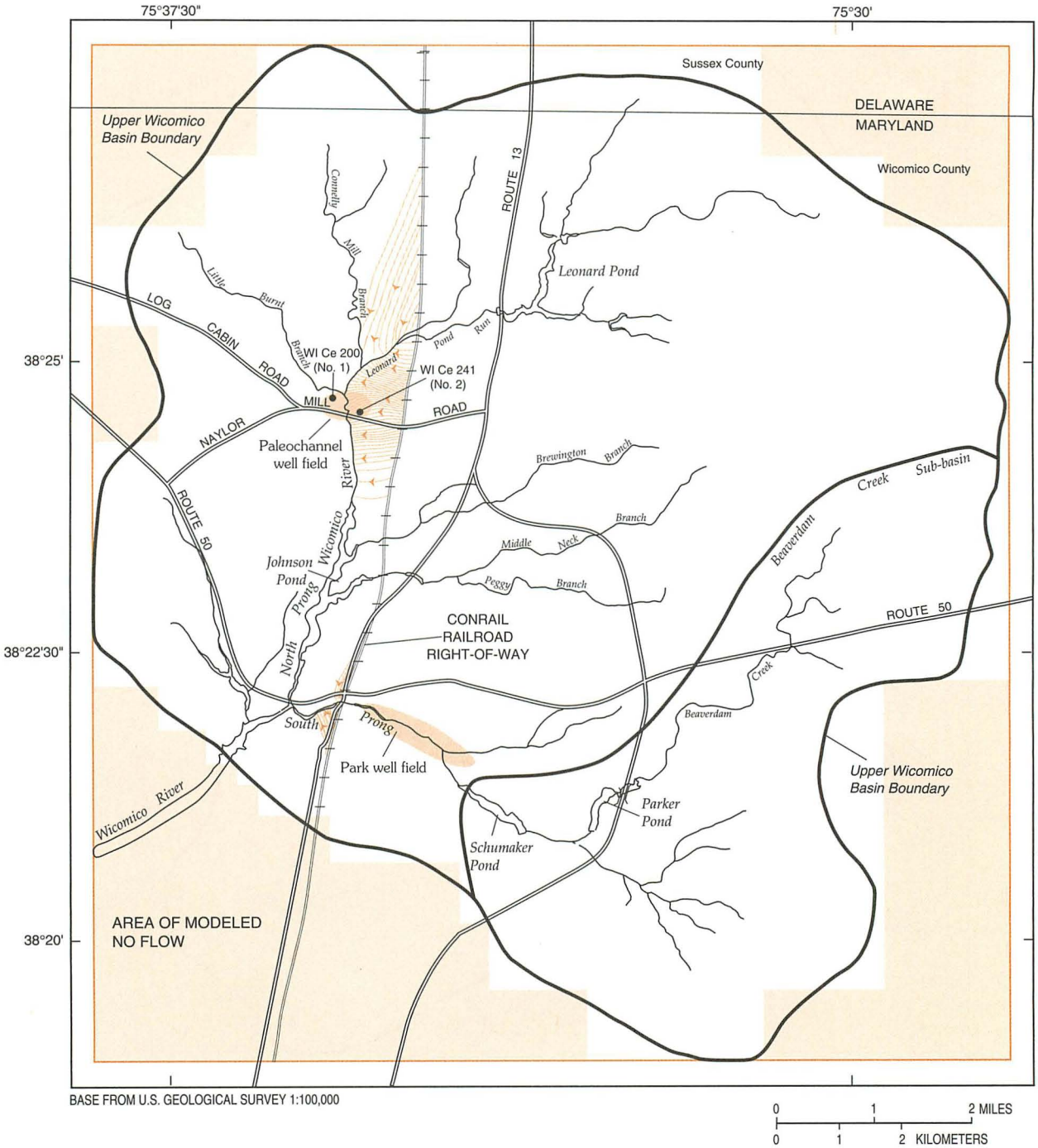
Figure 39. - Contributing areas for Park wells based on 1993 pumping conditions and a 50 percent increase in 1993 pumpage.



EXPLANATION

GROUND-WATER PARTICLE PATHS PROJECTED ONTO MODEL LAYER 1. ARROWS INDICATE THE FLOW DIRECTION.

Figure 40. - Forward particle tracks from the Wicomico County Newland Park landfill and the proposed Route 50 bypass.



EXPLANATION

GROUND-WATER PARTICLE PATHS PROJECTED ONTO MODEL LAYER 1. ARROWS INDICATE DIRECTION OF FLOW

Figure 41. - Forward particle tracks from Conrail railroad right-of-way.

Particles were tracked from five CERCLA or Superfund sites in the southern part of the basin (fig. 42). Particles from all five sites discharged to either Johnson Pond or the tidal part of the Wicomico River. No particles were intercepted by City of Salisbury wells.

Discharge of waste water (such as spray irrigation, overland flow systems, rapid infiltration systems, and injection wells) to ground water requires a ground-water discharge permit issued by the Maryland Department of the Environment (Maryland Department of the Environment, 1993). Particles were tracked forward from six sites with ground-water discharge permits that pose the greatest risk of reaching the city's well fields (fig. 43). In the simulation, well 2 was pumped at the 1993 rate. Previous simulations with well 1 pumping did not result in the capture of particles from these sites. Particles from sites 1, 2, 3, 4, and 6 discharged to streams. Particles from site 5, however, were intercepted by Paleochannel well 2. The shortest travel time from site 5 to well 2 was 11 years.

COMPARISON OF SIMULATED TRAVEL TIMES AND GROUND-WATER AGES

The estimated ages of ground water sampled in eight wells using tritium and CFC's (tab. 4) were compared to the range and median of travel times from the water table to the wells calculated using the ground-water-flow model (tab. 10). Particle travel times from wells to the water table were computed by MODPATH for twenty-five particles placed on the model-cell faces representing the wells. Wells WI Ce 13, 196, 204, 210, and 244 pumped relatively young water (less than 40 years) as determined by tritium and CFC analysis. The medians of simulated travel times of particles for these wells show good agreement with the tritium and CFC data. The tritium and CFC travel times for wells WI Ce 200 and Ce 241 indicate mixtures of older pre-bomb (Manokin aquifer?) water and younger Salisbury aquifer water. The median simulated travel time for water from well WI Ce 241 suggests a pre-bomb recharge date, whereas the median value for water from well WI Ce 200 suggests a more recent recharge date. The recharge age for well WI Ce 242 estimated by both methods is greater than 30 years.

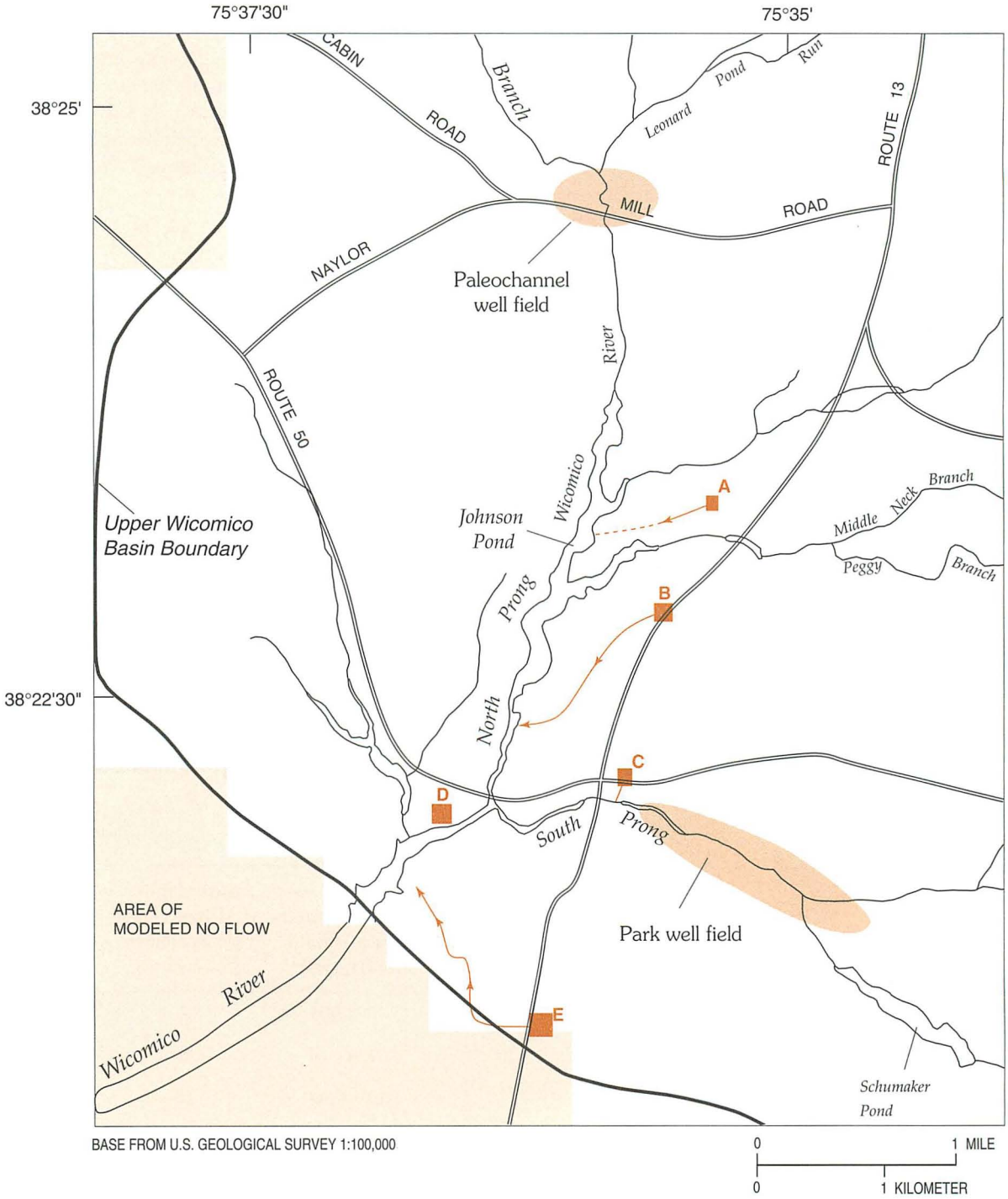
Table 10. - Comparison of simulated travel times and recharge age estimated using tritium and chlorofluorocarbons

Well number	Model cell (row, column, layer)	Recharge age [1] (years)	Range of flow-model travel times (water table to well) (years)[2]	Median flow-model travel time (water table to well) (years)[2]
WI Ce 13	67,39,2	7	5 - 2,000	8
WI Ce 196	21,20,1	10	7E-4 - 16	8
WI Ce 200	23,19,(3,4)	[3]	0.3 - 3,300	18
WI Ce 204	23,18,4	23	8 - 125	8
WI Ce 210	26,21,4	23	8 - 76	8
WI Ce 241	27,34,(3,4,5)	[3]	23 - 5,100	92
WI Ce 242	27,35,5	More than 30	190 - 2,800	270
WI Ce 244	66,40,2	2-24 or 32-39	0.1 - 5,300	17

[1] Age of water determined through a combination of tritium and chlorofluorocarbon analysis.

[2] Model simulation using even distribution of 1993 pumpage in Park well field and 1993 pumpage in Paleochannel well 1.

[3] Mixture of older pre-bomb (Manokin aquifer?) water and younger Salisbury aquifer water

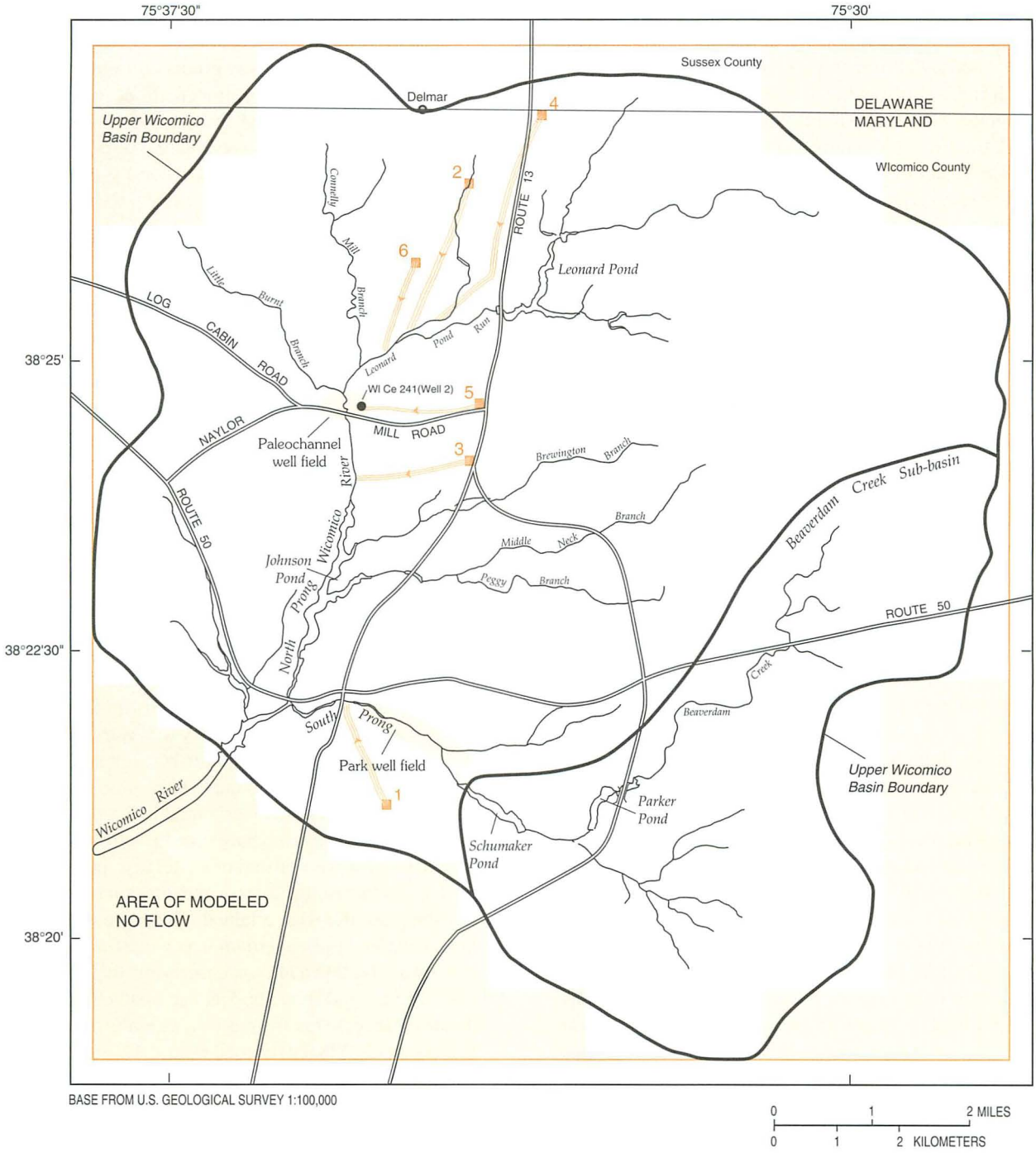


EXPLANATION

A — SUPERFUND SITE. LETTER NEXT TO SYMBOL IS IDENTIFIER.

GROUND - WATER PARTICLE PATHS PROJECTED ONTO MODEL LAYER 1. DASHED WHERE APPROXIMATE. ARROWS INDICATE DIRECTION OF FLOW.

Figure 42. - Forward particle tracks from Superfund sites.



EXPLANATION

3 SITE WITH GROUND - WATER DISCHARGE PERMIT. NUMBER IS IDENTIFIER.

GROUND-WATER PARTICLE PATHS PROJECTED ONTO MODEL LAYER 1. ARROWS INDICATE DIRECTION OF FLOW.

Figure 43. - Forward particle tracks from sites with ground-water discharge permits.

SUMMARY AND CONCLUSIONS

The City of Salisbury, in Wicomico County, Maryland, pumps ground water from two well fields (Paleochannel and Park well fields) that tap the Salisbury aquifer. In 1993, the Paleochannel well field pumped approximately 3.5 Mgal/d and the Park well field pumped approximately 1.7 Mgal/d. The Salisbury aquifer is a shallow, unconfined unit of high transmissivity and, as such, is vulnerable to contamination. To protect the well fields from contamination, it is necessary to determine what areas contribute recharge to the wells, and probable flow directions of potential ground-water contaminants. These determinations were the object of the study reported here. To accomplish this objective it was necessary to understand the hydrogeologic system and to incorporate that understanding into a ground-water-flow model. The ground-water-flow model, in conjunction with a particle-tracking routine, can then be used to estimate areas of recharge to the wells and direction of flow of potential ground-water contaminants.

The Salisbury aquifer is the water-table aquifer in the vicinity of the City of Salisbury. The aquifer's flow boundaries are coincident with the surface-water drainage divide of the Upper Wicomico River Basin. Water enters the aquifer as precipitation and as vertical flow from the deeper Manokin aquifer and flows toward streams and the tidal part of the Wicomico River. Water levels range in altitude from about 50 ft above sea level along the eastern edge of the basin to sea level at the outlet of the basin. Water levels in the Salisbury aquifer respond to seasonal changes in precipitation and to local pumping. Water levels fluctuate as much as 6 ft in some wells. Pumping in nearby wells results in greater water-level fluctuations, but this effect generally does not extend far from the pumping center. Pumping from the Salisbury aquifer at wells near streams can lower the amount of flow in the streams. For example, pumping from Paleochannel well 1 in 1993 lowered flow in Little Burnt Branch by 0.1 ft^3/s (about 45 gpm) near the well.

Transmissivity values of the Salisbury aquifer range from 8,200 ft^2/d to 57,000 ft^2/d, with an average of about 22,000 ft^2/d, in the Park well field. The Salisbury aquifer is separated from the underlying Manokin aquifer by a relatively thick (average about 60 ft) low permeability clay layer. This clay layer hydraulically separates the two aquifers, except in the northern part of the basin, where an erosional channel (paleochannel) has breached the confining layer. Sediments in the paleochannel consist of coarse-grained sands and gravels. Thickness of the Salisbury aquifer in the paleochannel approaches 200 ft. Transmissivity measured in the Paleochannel well field is as high as 53,500 ft^2/d.

Estimated recharge dates of water from wells screened in the Salisbury aquifer in and adjacent to the Park and Paleochannel well fields range from pre-1963 to 1992 on the basis of tritium and chlorofluorocarbon analysis. In general, the younger water was produced from the shallowest wells. The possible recharge dates of water from Paleochannel wells 1 and 2 were 1954-61, 1968-91 and 1954-61, 1975-present, respectively. The possible recharge dates for Park well 2 were 1954-61 and 1968-91.

Recharge from precipitation to the Salisbury aquifer was determined to be 13.4 in/yr. This value is equal to the median value of baseflow in Beaverdam Creek calculated over a 45-year period of continuous stream-flow record. Beaverdam Creek is located in the southeast part of the Upper Wicomico River Basin. This recharge value is consistent with other basins with similar characteristics located in central Delmarva Peninsula. Ground-water withdrawals from the Salisbury aquifer totaled 8.7 Mgal/d in 1993 within the Upper Wicomico River Basin. Average annual pumpage in 1993 was divided in the following way, (1) City of Salisbury—5.2 Mgal/d, (2) commercial and industrial—1.8 Mgal/d, (3) irrigation—1.0 Mgal/d, (4) domestic supply—0.7 Mgal/d, and (5) other public and private use—0.04 Mgal/d.

Ground-water flow within the Upper Wicomico River Basin was simulated using the U.S. Geological Survey's three-dimensional, finite-difference, ground-water-flow model (MODFLOW). The model included five layers to represent the Salisbury and Manokin aquifers, and the intervening confining bed. The clay (St. Marys Formation) at the base of the Manokin aquifer was treated as a no-flow boundary, as was the boundary of the ground-water divide in the Salisbury aquifer. Horizontal hydraulic conductivity in the top two layers, recharge, and river-bottom hydraulic conductivity were adjusted in an iterative procedure until a best fit between simulated and measured heads in the Salisbury aquifer were attained. The degree to which simulated baseflow in Beaverdam Creek matched the measured value was also taken into account during the calibration process. The ground-water budget calculated by the model indicates that recharge from precipitation accounts for about 85 percent (50 Mgal/d) of total inflow. The remaining inflow comes from constant-head cells (10 percent or 5.9 Mgal/d) and river leakage (5 percent or 2.8 Mgal/d). Total outflow was divided between river leakage (77 percent or 45 Mgal/d), flow to the tidal part of the Wicomico River (8 percent or 4.8 Mgal/d), and production wells (15 percent or 9 Mgal/d).

Contributing areas of the city's production wells were delineated using the ground-water-flow model and U.S. Geological Survey's particle-tracking routine (MODPATH). In the simulation in which Paleochannel well 1 is pumped at 1993 levels, two major contributing areas result within a 20-year time period; one extends outward from the well approximately 2 mi, and a second is disconnected from the

well and is approximately 1 mi long. In the simulation in which Paleochannel well 2 is pumped at 1993 levels, the 20-year contributing area extends eastward from the well approximately 2.2 mi. In both cases, the majority of the water entering the contributing area within a 20-year time period enters the upper and middle part of the well screens. The shapes of the contributing areas are irregular and are dependent on the values assigned for the model parameters (such as transmissivity, recharge, hydraulic conductivity, and porosity) during model calibration. Increasing recharge by 0.98 in/yr resulted in a narrowing and northern shift in the northeast-trending contributing area for the simulation of Paleochannel well 1 under 1993 pumping conditions. In addition, the northeast-trending portion was extended approximately 500 ft. Varying porosity values also affected the shape and extent of the contributing areas. A 10-percent reduction and increase in aquifer porosity lengthened and shortened, respectively, the contributing areas of Paleochannel well 1 by approximately 500 ft.

The contributing areas for the Park wells under 1993 pumping conditions extend approximately 10,000 ft upgradient from the pumped wells. Only three wells--wells 2a, 6a, and 8a--capture water from both sides of Park Pond. Curvature in the contributing areas reflects the configuration of the water-table surface, whereas the width is controlled by the amount of flow captured by the wells relative to the amount flowing to Park Pond. Water entering the ground-water system in the areas between the contributing areas discharges to Park Pond.

Increasing the pumping rate in the Paleochannel and Park well fields by 15 and 50 percent, respectively, causes the contributing areas to expand on average about 200 to 1,000 ft.

Particles tracked forward from sites with the potential to be sources of contamination to the Salisbury aquifer indicated that flow from two sites--the Conrail railroad right-of-way and a site with a ground-water-discharge permit--are captured by Paleochannel well 2 after 2 and 11 years of travel time, respectively. Particles tracked from the proposed U.S. Route 50 Bypass (located 0.3 mi south of the Paleochannel well field) flowed in a southerly direction away from the Paleochannel well field and discharged to the North Prong Wicomico River.

A comparison of recharge dates to particle travel times calculated by MODPATH helps to confirm the accuracy of the ground-water-flow model. Wells WI Ce 13, 196, 204, 210, and 244 pumped relatively young water (less than 40 years) as determined by tritium and CFC analysis. The medians of simulated travel times of particles for these wells show good agreement with these values. The tritium and CFC travel times for WI Ce 200 (Paleochannel well 1) and WI Ce 241 (Paleochannel well 2) indicate a mixture of older, pre-1954 Manokin aquifer(?) water and younger Salisbury aquifer water. The median simulated travel time for water from well WI Ce 241 suggests a pre-1954 recharge date, whereas the median value for water from well WI Ce 200 suggests a more recent recharge date (1975). The recharge age for well WI Ce 242, located 30 feet from well WI Ce 241 and screened in the lower part of the paleochannel, estimated by both methods is greater than 30 years.

SELECTED REFERENCES

Andres, A.S., 1986, Stratigraphy and depositional history of the post-Choptank Chesapeake Group: Delaware Geological Survey Report of Investigations No. 42, 39 p.

Bachman, L.J., 1984, Nitrate in the Columbia aquifer, central Delmarva Peninsula, Maryland: U.S. Geological Survey Water-Resources Investigations Report 84-4322, 51 p.

Bachman, L.J., and Wilson, J.M., 1984, The Columbia aquifer of the eastern shore of Maryland: Maryland Geological Survey Report of Investigations No. 40, 144 p.

Boggess, D.H., and Heidel, S.G., 1968, Water resources of the Salisbury area, Maryland: Maryland Geological Survey Report of Investigations No. 3, 69 p.

Busenberg, E., and Plummer, L.N., 1992, Use of chlorofluorocarbons (CCl_3F and CCl_2F_2) as hydrologic tracers and age-dating tools: The alluvium and terrace system of central Oklahoma: Water Resources Research, v. 28, no. 9, p. 2257-2283.

Busenberg, E., Weeks, E.P., Plummer, L.N., and Bartholomay, R.C., 1993, Age dating ground water by use of chlorofluorocarbons (CCl_3F and CCl_2F_2) and distribution of chlorofluorocarbons in the unsaturated zone, Snake River Plain aquifer, Idaho National Engineering Laboratory, Idaho: U.S. Geological Survey Water-Resources Investigations Report 93-4054, 47 p.

Chiang, W., and Kinzelbach, W., 1993, Processing MODFLOW: Hamburg, Germany, 92 p.

Coplen, T.B., 1993, Uses of environmental isotopes, *in* William M. Alley (editor), Regional ground-water quality: New York, N.Y., Van Nostrand Reinhold, p. 227-254.

Driscoll, F.G., 1986, Groundwater and wells, 2nd. edition: St. Paul, Minn., Johnson Division, 1,089 p.

Dunkle, S.A., Plummer, L.N., Busenberg, E., Phillips, P.J., Denver, J.M., Hamilton, P.A., Michel, R.L, and Coplen, T.B., 1993, Chlorofluorocarbons (CCL_3F and CCl_2F_2) as dating tools and hydrologic tracers in shallow groundwater of the Delmarva Peninsula, Atlantic Coastal Plain, United States: Water Resources Research, v. 29, no. 12, p. 3837-3860.

Freeze, R.A., and Cherry, J.A., 1979, Groundwater: Englewood Cliffs, N.J., Prentice-Hall, Inc., 604 p.

GeoSyntec Consultants, 1992, Phase II permit modification application report, Newland Park landfill, Wicomico County, Maryland, volume I: Timonium, Md., 95 p.

Hansen, H.J., 1966, Pleistocene stratigraphy of the Salisbury area, Maryland and its relationship to the lower Eastern Shore--a subsurface approach: Maryland Geological Survey Report of Investigations No. 2, 56 p.

____ 1981, Stratigraphic discussion in support of a major unconformity separating the Columbia Group from the underlying Upper Miocene aquifer complex in eastern Maryland: Southeastern Geology, v. 22, no. 3, p. 123-138.

Hansen, H.J., and Edwards, Jonathan, Jr., 1986, The lithology and distribution of pre-Cretaceous basement rocks beneath the Maryland Coastal Plain: Maryland Geological Survey Report of Investigations No. 44, 27 p.

Mack, F.K., and Thomas, W.O., 1972, Part 1--Channel-fill deposits near Salisbury, Maryland as determined by a 30-day pumping test: Maryland Geological Survey Bulletin 31, p. 1-60.

Maryland Department of the Environment, 1993, Ground water protection program annual report, July 1, 1992 to June 30, 1993: Baltimore, Md., Maryland Department of the Environment, Water Management Administration, 17 p.

McDonald, M.G., and Harbaugh, A.W., 1988, A modular three-dimensional finite-difference flow model: Techniques of Water-Resources Investigations of the U.S. Geological Survey, Book 6, Chap. A1, 586 p.

Meyer, R.R., and Bennett, R.R., 1955, A section on the Salisbury area, *in* Water resources of Somerset, Wicomico, and Worcester Counties: Maryland Department of Geology, Mines, and Water Resources Bulletin 16, 533 p.

Owens, J.P., and Denny, C.S., 1979a, Geologic map of Wicomico County: Maryland Geological Survey, 1 sheet, scale 1:62,500.

____ 1979b, Upper Cenozoic deposits of the central Delmarva Peninsula, Maryland and Delaware: U.S. Geological Survey Professional Paper 1067-A, 28 p.

Pazzaglia, F.J., 1993, Stratigraphy, petrography, and correlation of late Cenozoic middle Atlantic Coastal Plain deposits--Implications for late-stage passive-margin geologic evolution: Geological Society of America Bulletin, v. 105, p. 1617-1634.

Pettyjohn, W.A., and Henning, R., 1979, Preliminary estimates of ground-water recharge rates, related streamflow and water quality in Ohio: Ohio State University Water Resources Center Project Completion Report No. 522, 323 p.

Plummer, L.N., Michel, R.L., Thurman, E.M., and Glynn, P.D., 1993, Environmental tracers for age dating young ground water, *in* William M. Alley (editor), Regional ground-water quality: New York, N.Y., Van Nostrand Reinhold, p. 255-294.

Pollock, D.W., 1989, Documentation of computer programs to compute and display pathlines using results from the U.S. Geological Survey modular three-dimensional finite-difference ground-water flow model: U.S. Geological Survey Open-File Report 89-381, 188 p.

Rasmussen, W.C., and Andreasen, G.E., 1959, Hydrologic budget of the Beaverdam Creek Basin, Maryland: U.S. Geological Survey Water-Supply Paper 1472, 106 p.

Rasmussen, W.C., and Slaughter, T.H., 1955, Groundwater resources *in* Water resources of Somerset, Wicomico, and Worcester Counties: Maryland Department of Geology, Mines, and Water Resources Bulletin 16, 533 p.

Resources Management Associates, Inc., 1977, Wicomico County comprehensive water and sewer plan— 1993 amendment: Towson, Md., 273 p.

Riley, T.E., Plummer, N.L., Phillips, P.J., and Busenberg, E., 1994, The use of simulation and multiple environmental tracers to quantify groundwater flow in a shallow aquifer: Water Resources Research, v. 30, no. 2, p. 421-433.

Riley, T.E., and Pollock, D.W., 1989, Factors affecting areas contributing recharge to wells in shallow aquifers: U.S. Geological Survey Water-Supply Paper 2412, 21 p.

____ 1995, Effect of seasonal and long-term changes in stress on sources of water to wells: U.S. Geological Survey Water-Supply Paper 2445, 25 p.

Robertson, W.D., and Cherry, J.A., 1989, Tritium as an indicator of recharge and dispersion in a groundwater system in central Ontario: Water Resources Research, v. 25, no. 6, p. 1097-1109.

Todd, D.K., 1980, Groundwater hydrology: New York, N.Y., John Wiley and Sons, Inc., 535 p.

U.S. Environmental Protection Agency, 1987, Guidelines for delineation of wellhead protection areas: Washington, D.C., U.S. Environmental Protection Agency, Office of Ground-Water Protection, 131 p.

U.S. Geological Survey, 1964, Surface water records of Maryland and Delaware: Towson, Md., U.S. Geologi-

cal Survey, Water Resources Division, 138 p.

Weigle, J.M., 1972, Part 2: Exploration and mapping of Salisbury paleochannel, Wicomico County, Maryland: Maryland Geological Survey Bulletin 31, p. 61-124.

Wheeler, J.C., and Wilde, F.D., 1989, Ground-water use in the Coastal Plain of Maryland, 1900-1980: U.S. Geological Survey Open-File Report 87-540, 173 p.

White, K.E., and Sloto, R.A., 1990, Base-flow-frequency characteristics of selected Pennsylvania streams: U.S. Geological Survey Water-Resources Investigations Report 90-4160, 67 p.

Winter, T.C., LaBaugh, J.W., and Rosenberry, D.O., 1988, The design and use of a hydraulic potentiomanometer for direct measurements of differences in hydraulic head between groundwater and surface water: Limnology and Oceanography, v. 33, no. 5, p. 1209-1214.

Wolff, R.G., 1970, Field and laboratory determination of the hydraulic diffusivity of a confining bed: Water Resources Research, v. 6, no. 1, p. 194-203.

Zimmerman, H.B., Shackleton, N.J., Bachman, J., Kent, D.V., Baldauf, J.G., Kaltenback, A.J., and Morton, A.C., 1984, History of Plio-Pleistocene climate in the northeastern Atlantic deep sea drilling project hole 552A, *in* Roberts, D.G., Schnitker, D., and others, eds., Initial reports of the deep sea drilling project, vol. LXXXI: Washington, D.C., U.S. Government Printing Office, p. 861-875.

APPENDIXES

Appendix A. - Selected well records

[ft, foot; in., inch; m-d-yr, month-day-year; gal/min, gallons per minute; USGS, U.S. Geological Survey; --, not reported]

Well number	State permit number	Owner or location	Driller	Completion year	Altitude of land surface (ft)	Depth of well (ft)	Diameter (in.)		Depth to top of screen (ft)	Length of screen (ft)	Aquifer	Water level (ft below land surface)		Date measured (m-d-yr)	Yield (gal/min)	Hours pumped	Specific capacity ([gal/min]/ft)	Well number
							Casing	Screen				Static	Pumping					
WI Be 22	--	Avery, P.A.	P.A. Avery	1958	50	45	2	2	40	--	Salisbury	8.16	8-14-58	--	--	--	--	WI Be 22
WI Be 52	WI-88-0641	USGS	USGS	1989	47	49	2	2	46	--	Salisbury	--	--	--	--	--	--	WI Be 52
WI Be 56	WI-67-0036	Culver, L.G. and Sons	Ideal Well Drillers	1967	35	100	8	8	40	60	Salisbury	9.5	27	8-22-67	650	8	37	WI Be 56
WI Be 57	WI-92-1130	Culver, Thomas	Lifetime Well Drilling Co.	1993	42	100	12	12	20	80	Salisbury	--	--	--	--	--	--	WI Be 57
WI Be 58	WI-81-0791	Ellis, Robert	Lifetime Well Drilling Co.	1984	46	112	12	12	32	80	Salisbury	6	30	2-27-84	1,000	5	41.7	WI Be 58
WI Be 59	--	Lankford, Allan	--	--	40	150	--	--	--	--	Salisbury	--	--	--	--	--	--	WI Be 59
WI Bf 6	--	Lowe Bros.	--	--	49	27	1.25	1.25	--	--	Salisbury	9.20	--	9-23-47		--	--	WI Bf 6
WI Bf 20	31112	Edgewood Pipe and Block Co.	C. Cannon	1958	47	80	22	22	40	40	Salisbury	16	30	8-1-58	700	2	50	WI Bf 20
WI Bf 21	43937	Interstate Amiesite	E. R. Kauffman	1961	47	88	6	6	58	30	Salisbury	14	22	8-11-61	80	7	10	WI Bf 21
WI Bf 44	WI-65-0067	Shockley, Jr., Walter.	M. P. Brittingham	1965	48	99	2	2	86	13	Salisbury	16.9	29	1-26-65	20		2	WI Bf 44
WI Bf 64	WI-73-5254	State Line Motel	Ideal Well Drillers	1979	50	98	4	4	88	10	Salisbury	12	15	1-18-79	100	1	30	WI Bf 64
WI Bf 66	WI-73-3619	Tate Engineering Co.	Ideal Well Drillers	1977	50	55	4	4	50	5	Salisbury	10	16	6-22-77	70	1	12	WI Bf 66
WI Bf 69	--	USGS	Hardin Assoc.	1980	43	74	2	2	64	--	Salisbury	13.95	--	9-29-81	--	--	--	WI Bf 69
WI Bf 76	WI-73-7100	Town of Delmar	Delmarva Drilling Co., Inc.	1981	45	205	16	16	165	40	Salisbury	15.5	--	12-5-81	--	--	--	WI Bf 76
WI Bf 79	WI-73-4094	Wilber, Donald	Lifetime Well Drilling Co.	1977	42	87	8	8	47	40	Salisbury	9	40	11-29-77	490	7	16	WI Bf 79
WI Bg 3	--	Ward, H.	--	--	65	46	1.25	--	--	--	Salisbury	11.24	--	10-13-49	--	--	--	WI Bg 3
WI Ce 13	--	City of Salisbury	Shannahan Artesian Well Co.	1942	7	65	16	--	--	20	Salisbury	--	--	8- -42	200	--	--	WI Ce 13
WI Ce 16	--	City of Salisbury	Rulon	1943	7	47	4	--	--	--	Salisbury	3.05	--	7-23-47	--	--	--	WI Ce 16
WI Ce 87	--	Corona Nurseries	--	--	45	64	2	--	--	--	Salisbury	12.95	--	1-9-50	--	--	--	WI Ce 87
WI Ce 118	WI-02-8095	Md. National Guard	Shannahan Artesian Well Co.	1957	25	68	6	6	58	11	Salisbury	16.66	29.00	8-28-57	60	8	5	WI Ce 118

Appendix A. - Continued

Well number	State permit number	Owner or location	Driller	Completion year	Altitude of land surface (ft)	Depth of well (ft)	Diameter (in.)		Depth to top of screen (ft)	Length of screen (ft)	Aquifer	Water level (ft below land surface)		Date measured (m-d-yr)	Yield (gal/min)	Hours pumped	Specific capacity ([gal/min]/ft)	Well number
							Casing	Screen				Static	Pumping					
WI Ce 129	48650	Peninsula-Goslee, Inc.	P. W. Hubbert	1959	30	75	2	2	65	10	Salisbury	12	17	6-1-59	22	3	4.4	WI Ce 129
WI Ce 130	43538	Downing, Adele	E. P. Schultz	1961	22	72	2	2	62	10	Salisbury	13	28	7-8-61	50	2	3	WI Ce 130
WI Ce 148	54843	Mardel By-products Corp.	P.E.White	1964	42	197	8	8	182	15	Manokin	18.12	24.22	6-4-64	300	--	13.1	WI Ce 148
WI Ce 166	WI-65-0016	Wicomico Co. Elem. School	Ideal Well Drillers	1964	38	120	--	--	--	--	Salisbury	--	--	--	--	--	--	WI Ce 166
WI Ce 174	650054	Brown, Charles	Ideal Well Drillers	1964	32	54	2	2	44	10	Salisbury	16	31	10-29-64	30	1	2	WI Ce 174
WI Ce 184	16565	Messick Ice Co.	M. A. Pentz	1954	30	86	6	6	72	14	Salisbury	25	33	9-19-54	100	2	12	WI Ce 184
WI Ce 185	23894	Messick Ice Co.	M. A. Pentz	1957	30	83	8	8	72	11	Salisbury	23	69	6-17-57	500	4	11	WI Ce 185
WI Ce 195	--	USGS	USGS	1967	16	8	2	1.25	7	1	Salisbury	3.29	--	9-18-67	--	--	--	WI Ce 195
WI Ce 196	--	USGS	USGS	1967	16	7	1.25	1.25	6	1	Salisbury	0.13	--	9-18-67	--	--	--	WI Ce 196
WI Ce 197		USGS	USGS	1967	16	7	1.25	1.25	6	1	Salisbury	2.76	--	9-18-67	--	--	--	WI Ce 197
WI Ce 200	WI-67-0195	City of Salisbury, Paleochannel well 1	Layne Atlantic Co.	1967	29	163	26	16	80	80	Salisbury	10	34	10-18-67	4,000	720	200	WI Ce 200
WI Ce 204	WI-67-0191	City of Salisbury	Layne Atlantic Co.	1967	30	119	8	3	109	10	Salisbury	9.42	12.86	4-27-67	162	--	47.1	WI Ce 204
WI Ce 205	--	USGS	USGS	1967	18	7	2	1.25	6	1	Salisbury	--	--	--	--	--	--	WI Ce 205
WI Ce 206	--	USGS	USGS	1967	15	7	1.25	1.25	6	1	Salisbury	--	--	--	--	--	--	WI Ce 206
WI Ce 207	--	USGS	USGS	1967	16	7	2	1.25	6	1	Salisbury	--	--	--	--	--	--	WI Ce 207
WI Ce 208	--	USGS	USGS	1967	20	7	2	1.25	6	1	Salisbury	--	--	--	--	--	--	WI Ce 208
WI Ce 210	--	USGS	Ideal Well Drillers	1967	39	132	4	2	122	10	Salisbury	21.52	--	9-18-67	--	--	--	WI Ce 210
WI Ce 211	--	USGS	Ideal Well Drillers	1967	23	122	4	2	112	10	Salisbury	3.68	--	9-18-67	--	--	--	WI Ce 211
WI Ce 212-A	WI-67-0287	Deer's Head Realty Corp.	Ideal Well Drillers	1967	36	200	4	2	165	10	Manokin	18.70	--	11-17-67	--	--	6	WI Ce 212-A
WI Ce 213-A	WI-67-0288	J. W. Brittingham Estate	Ideal Well Drillers	1967	40	157	4	2	147	10	Salisbury	20.25	--	9-18-67	--	--	--	WI Ce 213-A
WI Ce 232	WI-73-5430	Wicomico Co. Dept. of Public Works	Delmarva Drilling Co., Inc.	1979	40	66	4	4	56	10	Salisbury	3	8	7-19-79	60	2	12	WI Ce 232
WI Ce 233	WI-73-1469	Wallace, Drusilla	E. R. Kauffman	1975	25	69	4	4	64	5	Salisbury	16	25	1-30-75	30	2	3	WI Ce 233

Appendix A. - Continued

Well number	State permit number	Owner or location	Driller	Completion year	Altitude of land surface (ft)	Depth of well (ft)	Diameter (in.)		Depth to top of screen (ft)	Length of screen (ft)	Aquifer	Water level (ft below land surface)		Date measured (m-d-yr)	Yield (gal/min)	Hours pumped	Specific capacity ([gal/min]/ft)	Well number
							Casing	Screen				Static	Pumping					
WI Ce 241	WI-88-0251	City of Salisbury, Paleochannel well 2	Delmarva Drilling Co., Inc.	1989	38	195	26	16	105	90	Salisbury	21	37	9-03-89	2,520	24	160	WI Ce 241
WI Ce 242	WI-81-4702	City of Salisbury	Delmarva Drilling Co., Inc.	1988	32	195	6	3	175	10	Salisbury	16	53	5-23-88	100	8	3	WI Ce 242
WI Ce 243	WI-81-2047	Wicomico County	Delmarva Drilling Co., Inc.	1985	40	100	8	8	20	80	Salisbury	20	35	10-08-95	450	4	30	WI Ce 243
WI Ce 244	WI-73-0382	City of Salisbury, Park well 2a	Ideal Well Drillers	1973	6.5	54	18, 12	12	38	16	Salisbury	6	24.6	8-15-73	418	24	20	WI Ce 244
WI Ce 245	WI-81-2384	City of Salisbury, Park well 5a	Layne Atlantic Co.	1986	6	62	18	12	29	26	Salisbury	4.3	23	3-12-86	521	12	25	WI Ce 245
WI Ce 246	WI-81-0603	City of Salisbury, Park well 6a	Delmarva Drilling Co., Inc.	1983	9	61	12	12	41	20	Salisbury	8.3	29	8-17-83	1,000	12	50	WI Ce 246
WI Ce 247	WI-81-2385	City of Salisbury, Park well 7b	Layne Atlantic Co.	1986	9	68	12	12	--	--	Salisbury	12	19.5	9-18-86	400	4	53	WI Ce 247
WI Ce 248	WI-81-0602	City of Salisbury, Park well 8a	Delmarva Drilling Co., Inc.	1983	10	64	12	12	44	20	Salisbury	9.95	42.2	10-04-83	1,016	12	31.5	WI Ce 248
WI Ce 249	--	City of Salisbury	--	--	9	20	4	--	--	--	Salisbury	6.85	--	7-22-93	--	--	--	WI Ce 249
WI Ce 250	WI-92-1340	Americhem, Inc.	Lifetime Well Drilling Co.	1993	35	90	4	4	50	40	Salisbury	15	20	1-13-93	100	5	20	WI Ce 250
WI Ce 251	WI-73-7085	Tolan Water Service, Inc.	Coastal Drilling Inc.	1981	36	80	4	4	70	10	Salisbury	22	27	10-22-81	30	1	6	WI Ce 251
WI Ce 252	WI-92-0218	Davis, Sr., Wilson	John D. Hines	1992	32	25	4	4	10	15	Salisbury	1	--	3-19-92	1	1	--	WI Ce 252
WI Ce 253	WI-81-0405	Naylor Mill Village Trailer Park	Dashiell Well Drilling	1983	43	150	4	4	140	10	Salisbury	15	21	5-24-83	75	1	12	WI Ce 253
WI Ce 254	WI-81-0404	Naylor Mill Village Trailer Park	Dashiell Well Drilling	1983	43	150	4	4	140	10	Salisbury	15	21	5-18-83	75	1	12	WI Ce 254
WI Ce 255	WI-81-2256	Cedarhurst Trailer Park	Larsons Well Drilling	1986	50	95	4	4	90	5	Salisbury	15	18	1-10-86	90	1	30	WI Ce 255
WI Ce 256	--	Braham, F. L.	--	--	40	24.5	2	--	--	--	Salisbury	22.27	--	10-05-93	--	--	--	WI Ce 256
WI Ce 259	WI-92-0954	Pocahontas Concrete, Inc.	Lifetime Well Drilling Co.	1993	15	60	8	8	40	20	Salisbury	10	15	1-29-93	500	2	100	WI Ce 259
WI Ce 260	WI-88-2488	Webcraft Technologies, Inc.	Lifetime Well Drilling Co.	1991	36	84	8	8	64	20	Salisbury	22	60	4-16-91	150	5	4	WI Ce 260
WI Ce 261	WI-66-0016	Royal Quality Foods, Inc., well 1	D. J. Shannahan	1965	20	68	8	8	51	17	Salisbury	18	39	10-22-65	350	10	17	WI Ce 261

Appendix A. - Continued

Well number	State permit number	Owner or location	Driller	Completion year	Altitude of land surface (ft)	Depth of well (ft)	Diameter (in.)		Depth to top of screen (ft)	Length of screen (ft)	Aquifer	Water level (ft below land surface)		Date measured (m-d-yr)	Yield (gal/min)	Hours pumped	Specific capacity ([gal/min]/ft)	Well number
							Casing	Screen				Static	Pumping					
WI Ce 262	WI-73-4817	Royal Quality Foods, Inc., well 2	S. V. Shannahan	1978	20	66	10	10	45	21	Salisbury	16	38	10-23-78	400	6	18	WI Ce 262
WI Ce 263	WI-70-0142	Royal Quality Foods, Inc., well 5	S. V. Shannahan	1971	20	70	12	10	46	24	Salisbury	14	38	5-21-71	400	24	17	WI Ce 263
WI Cf 3	--	Salisbury-Wicomico Co. Regional Airport	Rulon	1942	45	109	16	--	90	20	Salisbury	7	37	10-26-42	519	24	20	WI Cf 3
WI Cf 17	--	Methodist Church	--	--	45	41	1.25	1.25	--	--	Salisbury	7.62	--	9-23-47	--	--	--	WI Cf 17
WI Cf 52	--	Morris, T.	--	--	50	21	1.25	1.25	--	--	Salisbury	8.40	--	12-13-49	--	--	--	WI Cf 52
WI Cf 77	49744	City of Salisbury	Middletown Well Drilling Co.	1962	10	135	4	2	131	4	Manokin	+12.0	--	11-13-62	40	--	--	WI Cf 77
WI Cf 78	49745	City of Salisbury	Middletown Well Drilling Co.	1962	8	135	4	2	131	4	Manokin	+13.95	--	11-29-62	1.4	--	--	WI Cf 78
WI Cf 81	46375	Holiday Inn	Shannahan Artesian Well Co.	1962	40	96	6	5	86	10	Salisbury	12	14	4-11-62	36	4	18	WI Cf 81
WI Cf 95	28138	American Stores Co.	Sam Shannahan	1957	38	58	6	6	48	10	Salisbury	11	32	9-18-57	100	6	4.8	WI Cf 95
WI Cf 102	49156	Morris, N. L.	E. R. Kauffman	1962	44	74	4	4	44	30	Salisbury	13	25	10-9-62	40	2	3	WI Cf 102
WI Cf 104	29948	Beaver Run Elementary School	Shannahan Artesian Well Co.	1958	50	83	6	6	72	11	Salisbury	8.25	15	6-6-58	72	4	11	WI Cf 104
WI Cf 105	48318	Brittingham, J. W.	E. R. Kauffman	1962	45	84	6	6	10	74	Salisbury	12	28	8-10-62	300	4	20	WI Cf 105
WI Cf 107	--	Palone, T. S.	--	1963	42	38	1.5	--	--	--	Salisbury	13.86	--	9-18-63	--	--	--	WI Cf 107
WI Cf 112	30768	Rayner Brothers	Delmarva Drilling Co.	1958	55	100	22	22	60	40	Salisbury	3	39	6-3-58	600-1,000	3.5	--	WI Cf 112
WI Cf 117	38218	Twilley, M. W.	P. W. Hubbert	1960	33	77	2	2	70	7	Salisbury	14	23	4-13-60	75	2	8.3	WI Cf 117
WI Cf 147	--	W. F. Allen Co.	Middletown Well Drilling Co.	1964	41	80	2	2	60	20	Salisbury	18.5	--	11-11-64	--	--	--	WI Cf 147
WI Cf 154	43072	Perdue Farms, Inc., well #1	Shannahan Artesian Well Co.	1961	47	86	10	10	64	22	Salisbury	9.5	21	7-17-61	455	8	40	WI Cf 154
WI Cf 155	52300	Perdue Farms, Inc., well #2	Shannahan Artesian Well Co.	1963	47	93	10	10	67	26	Salisbury	12	43	7-20-63	610	8	20	WI Cf 155
WI Cf 184	WI-73-5934	Maryland Geological Survey	Delmarva Drilling Co., Inc.	1980	47	282	--	--	--	--	--	--	--	--	--	--	--	WI Cf 184
WI Cf 187	WI-73-5007	Christian School of Salisbury	Ideal Well Drillers	1978	50	59	4	2	54	5	Salisbury	16	19	11-9-78	75	1	25	WI Cf 187

Appendix A. - Continued

Well number	State permit number	Owner or location	Driller	Completion year	Altitude of land surface (ft)	Depth of well (ft)	Diameter (in.)		Depth to top of screen (ft)	Length of screen (ft)	Aquifer	Water level (ft below land surface)		Date measured (m-d-yr)	Yield (gal/min)	Hours pumped	Specific capacity ([gal/min]/ft)	Well number
							Casing	Screen				Static	Pumping					
WI Cf 188	--	Planning, C. I.	--	--	50	22	1.25	--	--	--	Salisbury	10	--	2-18-72	--	--	--	WI Cf 188
WI Cf 189	--	Luffman, W. C.	--	--	35	104	18	--	--	--	Salisbury	14.7	--	9-18-80	--	--	--	WI Cf 189
WI Cf 193	WI-73-2356	City of Salisbury, Park well 14a	E. R. Kauffman	1976	11	72	14	12	52	20	Salisbury	11	34	1-15-76	1,000	2	40	WI Cf 193
WI Cf 194	WI-73-7219	City of Salisbury, Park well 10a	Delmarva Drilling Co., Inc.	1982	7	58	18	12	34	24	Salisbury	3.2	30.9	3-25-82	602	8	22	WI Cf 194
WI Cf 195	WI-81-2575	City of Salisbury, Park well 16a	Delmarva Drilling Co., Inc.	1986	16	62	18	12	42	20	Salisbury	17.7	39.8	5-13-86	503	24	22.8	WI Cf 195
WI Cf 196	WI-88-2802	Wicomico County Detention Center	Water Systems, Inc.	1991	39	80	4	4	70	10	Salisbury	20	24	7-22-91	60	2	15	WI Cf 196
WI Cf 197	WI-92-1919	Brittingham Plant Farms	E. D. Crump	1994	50	80	8	8	20	60	Salisbury	12	52	06-22-94	800	1	20	WI Cf 197
WI Cf 198	WI-70-0154	Townsend, Lee	Ideal Well Drillers	1970	45	78	6	6	58	20	Salisbury	21	50	6-3-70	200	4	7	WI Cf 198
WI Cf 199	WI-73-2788	Townsend, Lee	Lifetime Well Drilling Co.	1976	45	70	6	6	40	30	Salisbury	8	25	6-22-76	195	4	11	WI Cf 199
WI Cf 200	WI-73-0024	Perdue Farms, Inc., well #4	Shannahan Artesian Well Co.	1972	47	92	10	10	67	25	Salisbury	11	17	9-26-72	148	10	25	WI Cf 200
WI Cf 201	WI-93-0107	City of Salisbury, Park well #17	A. C. Schultes of Delaware	1994	35	90	14	14	40	50	Salisbury	29	56	11-1-94	1,280	12	47	WI Cf 201
WI Cf 202	WI-93-0106	City of Salisbury, Park well #18	A. C. Schultes of Delaware	1994	32	92	14	14	42	50	Salisbury	23	59	11-9-94	1,100	12	30	WI Cf 202
WI Cf 203	WI-88-0927	Stadler, Randy	Larson Wells, Inc.	1989	25	60	4	4	50	10	Salisbury	18	21	8-4-89	80	1	30	WI Cf 203
WI Cf 204	WI-73-2788	Townsend, Jeffery	Lifetime Well Drilling Co.	1976	45	70	6	6	40	30	Salisbury	8	25	6-22-76	195	4	10	WI Cf 204
WI Cf 205	--	Shockley, Albert	--	--	42	90	--	--	--	--	Salisbury	--	--	--	750	--	--	WI Cf 205
WI Cf 206	WI-81-4498	Wilber, Allan	Lifetime Well Drilling Co.	1988	50	96	8	8	16	80	Salisbury	13	14	4-24-88	75	5	75	WI Cf 206
WI Cf 207	WI-88-0498	Brown, Edward	Lifetime Well Drilling Co.	1989	42	80	8	8	70	10	Salisbury	18	19	3-2-89	70	5	70	WI Cf 207
WI Cg 22	--	Leonard, Morris J.	--	1920	60	34	1.25	--	--	--	Salisbury	9.80	--	7-13-50	--	--	--	WI Cg 22
WI Cg 26	--	State Highway Administration	USGS	1950	51	17	1	1	16	1	Salisbury	1.56	--	4-7-50	--	--	--	WI Cg 26

Appendix A. - Continued

Well number	State permit number	Owner or location	Driller	Completion year	Altitude of land surface (ft)	Depth of well (ft)	Diameter (in.)		Depth to top of screen (ft)	Length of screen (ft)	Aquifer	Water level (ft below land surface)		Date measured (m-d-yr)	Yield (gal/min)	Hours pumped	Specific capacity ([gal/min]/ft)	Well number
							Casing	Screen				Static	Pumping					
WI Df 47	--	USGS	USGS	1952	48	56	3	--	--	--	Salisbury	6.8	--	5-12-52	--	--	--	WI Df 47
WI Df 51	51798	Payne, B. T.	M. O. Ford	1963	50	74	2	2	66	8	Salisbury	7	--	5-25-63	14	2	--	WI Df 51
WI Df 60	WI-73-4552	Dauville, John R.	Dashiell Well Drilling	1978	45	90	8	4	80	10	Salisbury	3	10	5-16-78	35	1	5	WI Df 60
WI Df 62	WI-88-1795	Nutters Crossing Golf Course	A. C. Schultes of Delaware	1990	45	246	8	8	156	90	Manokin	20	80	7-19-90	275	10	5	WI Df 62
WI Df 63	WI-88-1433	Nutters Crossing Golf Course	A. C. Schultes of Delaware	1990	45	246	4	4	156	90	Manokin	20	54	·5-17-90	70	3	2	WI Df 63

Appendix B. - Lithologic description of river-bottom sediment in Park Pond, collected October 25, 1994

Sample 1A (hand-driven, split-spoon core; length = 6 inches):
Sand, fine to medium, coarser grains and small pebbles at base, dark gray, muddy with silty inclusions, and black organic laminae in first 4 inches; half-inch clayey, black, organic layer at 4 inches, light gray sand at base of core.

Sample 1B (hand-driven, split-spoon core; length = 9 inches):
Sand, silty; muddy layer top half-inch; light gray, well sorted sand to 6 inches and at base; black, organic clay layer between 6 and 8 inches.

Sample 1C (hand-driven, split-spoon core; length = 8 inches):
Sand, fine to coarse with small sub-rounded pebbles, light to dark gray, some silty, black organic inclusions; bottom inch consists of highly oxidized, orange, medium sand.

Sample 2A (hand-driven, split-spoon core; length = 7 inches):
Clay, silty, muddy, black, organic, with fine sand inclusions; fine to medium, gray sand at bottom of core with some gravel.

Sample 2B (hand-driven, split-spoon core; length = 8 inches):
Sand, fine to medium, light gray to black, with black, organic, silt inclusions and some small pebbles.

Sample 2C (hand-driven, split-spoon core; length = 9 inches):
Sand, fine to medium, light gray to black, with black, organic, silt inclusions.

Sample 3A (hand-driven, split-spoon core; length = 7 inches):
Sand, silty to fine, light to dark gray, with some silt inclusions; muddy, partially decomposed organic material at top two inches.

Sample 3B (hand-driven, split-spoon core; length = 7 inches):
Sand, fine to medium, light to dark gray, small, partially decomposed organic material at top two inches and at base.

Sample 3C (hand-driven, split-spoon core; length = 6 inches):
Sand, silty to medium, light to dark gray, with small pods and black, organic, clayey silt and partially decomposed organic material.

Sample 4 (2-inch hand auger):
Depth 0 - 3.5 feet: Sand, fine to coarse, light gray, with some partially decomposed organic material.

Depth 3.5 - 4 feet: Sand, fine to coarse, light to dark gray, with pods of sticky, black, organic clay.

Sample 5 (shovel sample: approximate depth between 0 and 2 feet)
Sand, fine to coarse, light gray, with some small pebbles.

Appendix C. - Water levels in wells screened in the Salisbury aquifer used in calibrating the ground-water-flow model and preparing a water-table elevation map

Well number	Model cell (row,column)	Model layer	Well screen position (feet related to sea level)	Water level (feet related to sea level)	Date measured (month/day/year)	Median water level (feet related to sea level)
WI Be 22	2, 50	1	8-3	39.84	8/14/58	39.84
WI Be 52	2, 4	1	1-(-2)	37.66	5/20/89	36.68
				36.95	6/12/89	
				36.41	11/8/89	
				37.34	4/4/90	
				34.27	7/25/90	
				33.54	8/6/91	
WI Be 56	4, 4	2	-5-(-65)	27.32	10/5/93	29.30
				31.28	6/5/94	
WI Bf 6	3, 78	1	?-22	40.18	9/23/47	40.18
WI Bf 20	3, 61	2	7-(-33)	30.09	8/15/63	30.09
WI Bf 21	3, 52	2	-11-(-41)	30.61	8/11/61	30.61
WI Bf 44	7, 76	2	-38-(-51)	31.10	5/27/59	31.10
WI Bf 64	1, 75	2	-38-(-48)	36.20	9/9/80	37.66
				37.60	10/8/80	
				40.70	6/8/81	
				37.40	9/29/81	
				37.71	10/1/93	
				42.14	5/6/94	
WI Bf 66	2, 75	1	0-(-5)	37.70	10/8/80	38.3
				39.50	6/8/81	
				37.68	9/29/81	
				38.95	4/15/82	
WI Bf 69	6, 74	2	-19-(-29)	31.05	9/29/81	31.05
WI Bg 3	4, 80	1	?-20	53.76	10/13/49	53.76
WI Ce 13	67, 39	2	?-(-48)			*3.0
WI Ce 16	71, 49	2	?-(-40)	3.84	7/22/93	4.52
				5.2	5/3/94	
WI Ce 87	50, 3	2	?-(-20)	32.05	1/9/50	32.05
WI Ce 118	56, 3	2	-32-(-44)	7.51	9/11/63	7.51
WI Ce 129	56, 37	2	-35-(-45)	18	6/-/59	18
WI Ce 130	55, 9	2	-40-(-50)	10.61	9/12/63	10.61
WI Ce 166	53, 4	2	?-(-80)	25.31	7/31/64	25.31
WI Ce 174	79, 49	2	-12-(-22)	10.50	10/29/64	10.50
WI Ce 195	22, 20	1	9-8	16.96	9/18/67	15.80
				14.48	9/29/93	
				15.80	5/3/94	
WI Ce 196	21, 20	1	10-9	16.62	9/18/67	15.68
				13.75	9/29/93	
				15.68	5/3/94	
WI Ce 197	19, 20	1	10-9	17.42	9/18/67	
				14.61	9/29/93	*15.7
				15.97	5/3/94	
WI Ce 204	23, 18	4	-81-(-85)			approximately 15.7
WI Ce 205	22, 15	1	12-11	18.60	9/18/67	17.99
				17.04	9/29/93	
				17.99	5/3/94	
WI Ce 206	23, 24	1	9-8	16.45	9/18/67	15.24
				14.20	10/5/93	
				15.24	5/3/94	
WI Ce 207	23, 23	1	10-9	15.73	9/18/67	15.06
				14.37	9/29/93	
				15.06	5/3/94	
WI Ce 208	17, 13	1	14-13	21.0	9/18/67	20.29
				18.89	15/5/93	
				20.29	5/3/94	
WI Ce 210	26, 21	4	-83-(-93)			*17
WI Ce 211	23, 15	4	-89-(-99)	21.43	9/18/67	17.60
				17.60	9/29/93	
				19.81	5/3/94	
WI Ce 213	25, 34	4	-107-(-117)			*18

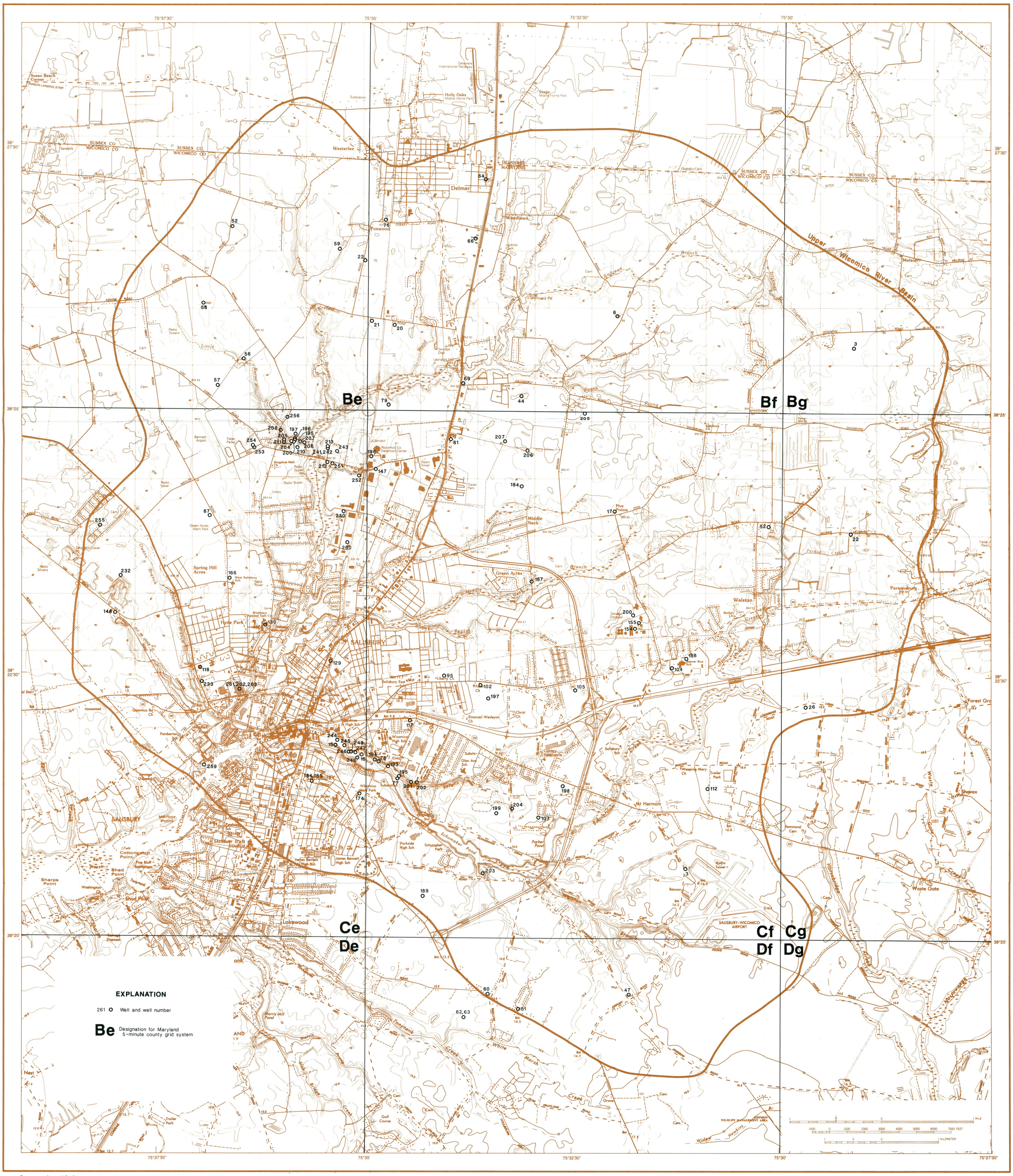
Appendix C. - Continued.

Well number	Model cell (row,column)	Model layer	Well screen position (feet related to sea level)	Water level (feet related to sea level)	Date measured (month/day/year)	Median water level (feet related to sea level)
WI Ce 232	53, 1	2	-16-(-26)	34.10	7/24/80	32.98
				31.92	10/9/80	
				34.05	6/10/81	
				30.80	9/30/81	
WI Ce 233	57, 3	2	-39-(-44)	6.50	7/16/80	6.50
				6.07	10/9/80	
				7.14	6/10/81	
WI Ce 242	27, 35	5	-143-(-153)	17.22	9/28/93	18.43
				19.65	5/3/94	
WI Ce 243	39, 35	1, 2	21-(-59)	20.25	7/6/93	20.25
WI Ce 246	70, 44	2	-32-(-52)	1.40	7/22/93	1.75
				1.75	9/29/93	
				2.67	5/3/94	
WI Ce 248	72, 48	2	-34-(-54)	4.09	5/3/94	4.09
WI Ce 249	70, 45	1	?-(-11)	2.16	7/22/93	2.48
				2.48	9/29/93	
				3.47	5/3/94	
WI Ce 250	50, 42	2	-15-(-55)	15.50	10/15/93	16.65
				17.80	5/6/94	
WI Ce 251	35, 37	2	-34-(-44)	15.60	10/15/93	16.88
				18.15	5/6/94	
WI Ce 252	40, 48	1	22-7	21.65	10/15/93	23.48
				25.30	5/6/94	
WI Ce 253	26, 6	4	-97-(-107)	22.03	10/5/94	23.90
				25.78	5/6/94	
WI Ce 254	25, 6	4	-97-(-107)	22.61	10/5/93	24.49
				26.38	5/6/94	
WI Ce 255	51, 1	2	-40-(-45)	32.80	10/5/93	36.10
				39.40	5/6/94	
WI Ce 256	11, 16	1	?-16	20.23	10/5/93	20.23
WI Cf 3	81, 78	2	-45-(-65)			*38.0
WI Cf 17	49, 78	1	?-4	37.38	9/23/47	37.38
WI Cf 52	51, 79	1	?-29	41.60	12/13/49	41.60
WI Cf 81	21, 73	2	-46-(-56)	24.93	9/12/63	24.93
WI Cf 95	57, 73	2	-10-(-20)	27.18	9/12/63	27.18
WI Cf 102	57, 75	1, 2	0-(-30)	29.01	9/16/63	29.01
WI Cf 104	56, 78	2	-22-(-33)	40.48	9/18/63	40.48
WI Cf 105	57, 77	1, 2	35-(-39)	35.96	9/18/63	35.96
WI Cf 107	80, 76	1	?-4	28.14	9/18/63	28.14
WI Cf 112	78, 78	2	-5-(-45)	48.49	9/18/63	48.49
WI Cf 117	61, 68	2	-37-(-44)	14.09	9/18/63	14.09
WI Cf 147	36, 55	2	-18-(-38)			*24.1
WI Cf 187	54, 76	1	-4-(-9)	34.95	7/10/80	34.95
WI Cf 188	56, 78	1	33-28	34.00	9/10/80	34.68
				33.50	10/8/80	
				35.35	6/8/81	
				37.40	9/30/81	
WI Cf 189	82, 71	2	?-(-69)	20.30	9/18/80	20.04
				19.40	10/8/80	
				20.15	6/10/81	
				19.94	10/1/93	
WI Cf 195	77, 65	2	-26-(-46)	5.05	7/22/93	5.07
				5.09	9/29/93	
WI Cf 196	31, 53	2	-31-(-41)	21.30	10/15/93	23.15
				25.00	5/6/94	
WI Cg 22	51, 80	1	?-27	50.20	7/13/50	50.20
WI Cg 26	58, 79	1	?-34	49.14	4/7/50	49.14
WI Df 47	83, 78	1	?-(-8)	40.70	5/12/52	40.70
WI Df 51	83, 76	2	-16-(-24)	42.69	8/15/63	42.69
WI Df 60	83, 75	2	-35-(-45)	32.12	10/1/80	32.18
				32.20	10/8/80	
				32.60	6/10/81	
				32.15	9/30/81	

* Median of long-term water-level record was used (figs. 8 and 9).

Appendix D. - Appropriated ground-water use in the Salisbury aquifer over 10,000 gallons per day in the Upper Wicomico River Basin

Owner or name	Maryland State ground-water appropriation permit number (GAP)	U.S. Geological Survey well number -- see appendix A (Owners well number)	Amount appropriated (1,000 gallons per day)		Average daily pumpage in 1993 (1,000 gallons per day)	Type of use
			Average	Maximum		
Perdue Farms	WI55G001	WI Cf 154 (well 1) WI Cf 155 (well 2) WI Cf 200 (well 4)	1,700	1,900	990	Industrial
Atlantic Ice Co.	WI56G001	WI Ce 184 WI Ce 185	65	700	580	Industrial
Royal Quality Foods, Inc.	WI66G007	WI Ce 261 (well 1) WI Ce 262 (well 2) -- (well 4) WI Ce 263 (well 5)	500	750	120 (7/93-12/93)	Industrial
Town of Delmar	WI81G008	WI Bf 76	210	310	0.34	Public supply
Walston Trailer Park	WI68G016	--	15	25	10	Public supply
City of Salisbury	WI73G001	**Paleochannel well field:** WI Ce 200 (well 1) WI Ce 241 (well 2) **Park well field** WI Ce 244 (well 2a) WI Ce 246 (well 6a) WI Ce 247 (well 7b) WI Ce 248 (well 8a) WI Cf 193 (well 14a) WI Cf 194 (well 10a) WI Cf 195 (well 16a) WI Cf 201 (well 17) WI Cf 202 (well 18)	7,600	10,000	5,200	Public supply
Arundel Concrete	WI73G003	WI Ce 259	24	50	5	Industrial
Cedarhurst Trailer Park	WI74G008	WI Ce 255	12	20	8	Public supply
Webcraft Technologies	WI81G016	WI Ce 260	125	160	120	Industrial
Louis E. Brown	WI89G004	WI Cf 207	95	570	--	Irrigation
Shockley Farm, Inc.	WI90G006	WI Cf 205	212	1,260	--	Irrigation
Oakwood Sod Farm, Inc. (Brown Rd.)	WI92G011	WI Cf 206	108	654	--	Irrigation
Oakwood Sod Farm, Inc. (Hampshire Rd.)	WI92G013	WI Bf 79	81	490	--	Irrigation
Randolf Stadler	WI92G014	WI Cf 203	16	32	--	Irrigation
Lee Townsend (Gunby Rd.)	WI93G004	WI Cf 199	54	324	--	Irrigation
Lee Townsend (Mt. Herman Rd.)	WI93G005	WI Cf 198	20	118	--	Irrigation
Lee Townsend (Gunby Rd.)	WI93G006	WI Cf 204	12	68	--	Irrigation
Lee Townsend	WI93G007	--	16	93	--	Irrigation
Alan Lankford	WI93G011	WI Be 59	150	900	--	Irrigation
Thomas Culver	WI93G012	WI Be 57	168	1,020	--	Irrigation
Brittingham Plant Farms	WI93G040	WI Cf 197	27	163	--	Irrigation



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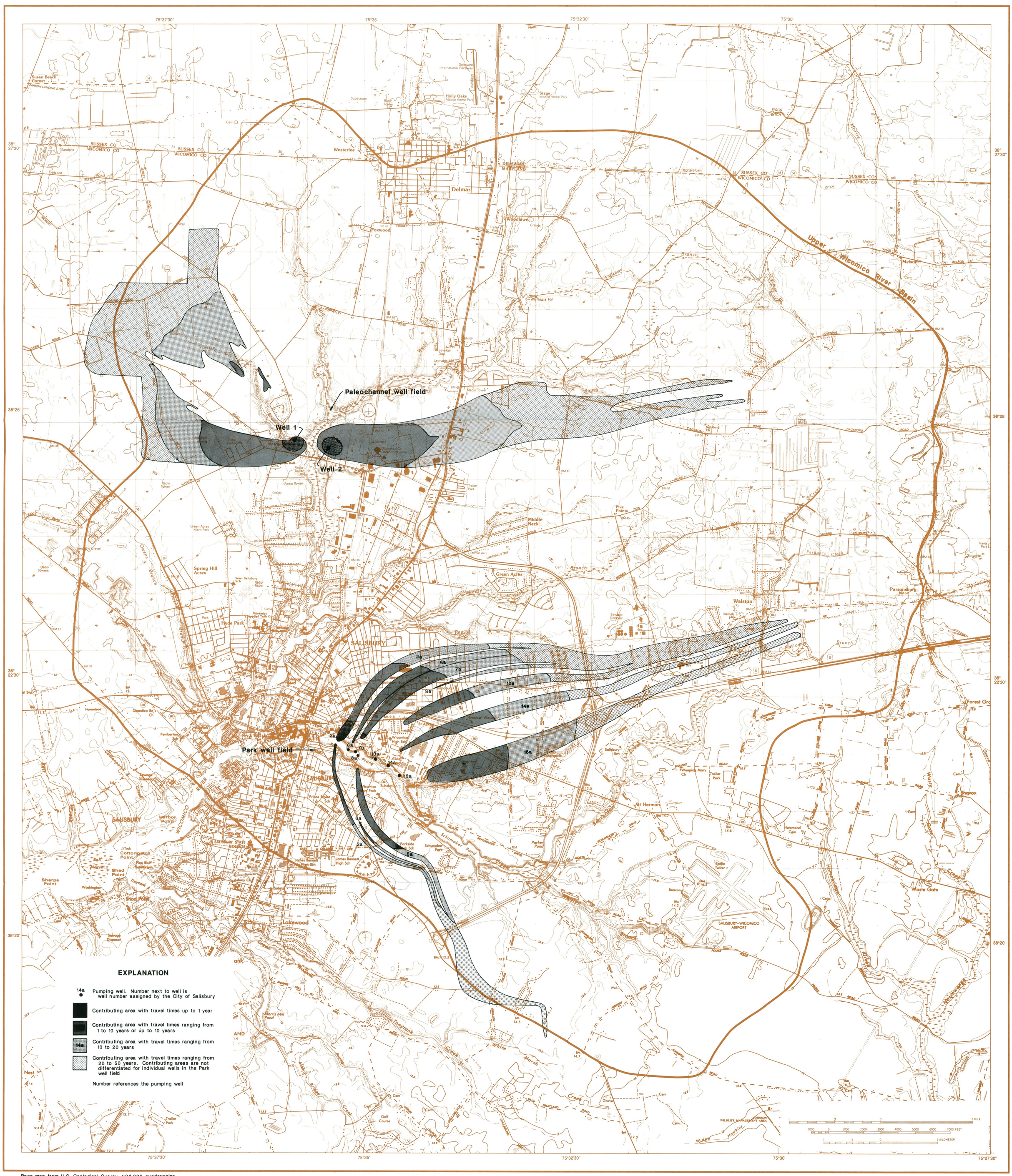

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